Petrogenesis of the East Fork Member Rhyolites, Valles Caldera, New Mexico, USA

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PETROGENESIS OF THE EAST FORK MEMBER RHYOLITES, VALLES CALDERA, NEW MEXICO, USA

by

Carla M. Eichler

Bachelor of Science
University of Arizona
2008

A thesis submitted in partial fulfillment of the requirements for the

Master of Science in Geoscience

Department of Geoscience
College of Science
Graduate College

University of Nevada, Las Vegas
December 2012
THE GRADUATE COLLEGE

We recommend the thesis prepared under our supervision by
Carla Eichler

entitled
Petrogenesis of the East Fork Member Rhyolites, Valles Caldera, New Mexico, USA

be accepted in partial fulfillment of the requirements for the degree of

Master of Science in Geoscience
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December 2012
ABSTRACT

Petrogenesis of the East Fork Member Rhyolites, Valles Caldera, New Mexico, USA

by

Carla M. Eichler

Dr. Terry Spell, Examination Committee Chair
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The most recent volcanism in the Valles caldera is represented by the El Cajete Pyroclastic Beds (ECPB), Battleship Rock Ignimbrite (BRI), and Banco Bonito Flow (BBF) as well as the VC-1 rhyolite, which are collectively known as the East Fork Member (EFM) of the Valles Rhyolite. The EFM was erupted at approximately 55 ka and 40 ka after an approximate 460 ka lull in volcanism. Previous studies suggested a mafic intrusion at depth triggered the eruptions. This thesis represents the first detailed study of the EFM.

Crystal assemblages consist of plagioclase, biotite, clinopyroxene, orthopyroxene, amphibole, sanidine, quartz, and oxides. Electron probe microanalysis and detailed petrography indicates that two distinct crystal populations are present in the ECPB, BRI, and BBF. Large (≥ 1 mm), typically resorbed or subhedral crystals represent one population, and small (≤ 0.5 mm), generally euhedral crystals represent the other. The large resorbed plagioclase crystals typically have rim overgrowths. Both normal and reverse zonation is present. $^{40}$Ar/$^{39}$Ar geochronology performed on euhedral biotite and groundmass glass from the BBF returned isochron ages of $478 \pm 27$ ka and $575 \pm 15$ ka
and total gas ages of 125 ± 1 ka and 129.82 ± 0.80 ka, respectively. High Mg numbers of large euhedral biotite and \(^{40}\text{Ar}/^{39}\text{Ar}\) ages older than the accepted age range indicate these crystals are xenocrystic. Radiogenic isotopes are consistent with mixing between the mantle and lower crustal reservoirs. General trends are evident between whole-rock major and trace elements with increasing SiO\(_2\). In general, incompatible trace elements increase and compatible trace elements decrease. Incompatible trace element ratios indicate the presence of a single magma batch.

The heterogeneity in crystal morphology and chemistry can be explained by a model in which partial melting of mid- to deep continental crust occurred due to an intrusion of an intermediate composition magma. Magma mixing and an exchange of crystals took place between the partial melt and the intruding magma. The hybrid magma rose to the upper crust. Trends in the trace element data indicates fractional crystallization was the last process to take prior to eruption. The geochemical and isotopic data from this study are best explained by a modified version of the rapid production and eruption model put forth by Huppert and Sparks (1988).
ACKNOWLEDGEMENTS

First and foremost, I would like to thank my advisor Dr. Terry Spell for his guidance, patience, and support throughout this project. His constructive critiques challenged my thinking and substantially improved this finished thesis. His efforts have been greatly appreciated. I would also like to thank my committee members Drs. Adam Simon, Michael Wells, George Rhee, John Wolff, and Barbara Luke for their valuable critique, suggestions, and support. I would also like to thank Dr. Sean Mulcahy for his assistance with the electron microprobe analyses, and Kathy Zanetti for her assistance with the argon mass spectrometer. I thank Dr. G. Lang Farmer and the staff of the TIMS lab at the University of Colorado-Boulder for their assistance with the TIMS analyses. I also thank Dr. John Wolff and the staff of the GeoAnalytical lab at Washington State University for their assistance with the XRF and ICPMS analyses.

To my family and friends who helped me overcome challenges and supported me during life’s troubles, I offer my most sincere thanks. I am indebted to you all for your love. I would especially like to thank Jon Baker and Chris Byrnes for their unconditional support, encouragement, and friendship. Their brilliant logic and laughter has helped me beyond measure.

This thesis was funded by the Nevada Isotope and Geochronology Laboratory, a Geological Society of America Student Research grant, and the Bernada French, Edwards-Olswang Geology and Nate Stout Memorial scholarships. I sincerely thank all donors for their financial support.

This thesis is dedicated to my father, Kevin Eichler, who encouraged me to pursue a higher education. His love and support will never be forgotten.
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CHAPTER 1

INTRODUCTION

Although large continental silicic magma systems have been studied for a number of decades, there remains debate as to their origin and the processes that take place before and after large-scale caldera-forming eruptions. Products of the pre-caldera eruptions are usually destroyed by the caldera-forming eruption. However, the post-collapse eruptive units are often preserved and thus can therefore offer insight into the processes that take place in caldera-forming systems. Several models have been put forth to explain the magma production, evolution, and longevity of these systems.

1.1 Previous Models

Early research on granites and rhyolites generally considered two different end-member models for the origin and longevity of silicic magma systems in the crust. The first model is that of a large mid- to upper-crustal convecting magma body which is thermally sustained by underlying mafic sills (e.g., Smith, 1979; Hildreth, 1981; Lipman, 1984). In this model, such magma bodies are thermally stable and may remain in a liquid state in the crust for timescales greater than 1 Ma. Periodic eruptions that take place could range from large-scale explosive to small-scale effusive events. These eruptions all tap the same magma chamber, which is thermally and compositionally zoned. Smith (1979) suggests that Ta, U, Th, Rb, Cs, Li, F, Cl, Pb, Zn, Be, Sn, W, Mo, B, Sm, and the HREEs are concentrated upwards in the magma chamber (as seen in the erupted units) whereas Ba, Sr, Eu, Ti, Cr, Co, Sc, Au, and Cu are concentrated downwards. Additionally, Hildreth (1981) suggests that Ca, Al, Fe, Mg, Ti, and the LREEs are
concentrated downwards in the magma chamber. Hildreth (1981) accounts for some compositional zoning by diffusion of elements across boundary layers. Mafic magma, which ponds at the base of the system, provides thermal input needed for the magma in the chamber to convect. The main body of magma is composed of three parts: the quasi-stagnant upper layer of the most fractionated material, a dominant convecting middle layer, and the lowest layer which is in contact with the mafic magma. Depending upon the size of the magma chamber and eruption, a portion of the mafic magma ponded at the base of the chamber may be erupted.

The other end-member model suggests that the silicic magmas are generated on relatively rapid timescales by the injection of basalt into the crust causing it to partially melt (Huppert and Sparks, 1988). A sequence of events leads up to the eruption of rhyolite. First, basalt is injected into the lower crust. Basaltic sills provide an ideal heating surface since they spread laterally and can melt a large area of crust immediately above them. The lower continental crust is hot and, in some cases, may be layered. Thus, it provides an ideal location for sills to form and melt the crust. After the emplacement of the sill, newly formed silicic magma begins to collect and convect within days. These silicic magmas may rapidly rise through the crust and erupt intermittently as rhyolites at the surface. Alternatively, they may coalesce in the shallow crust and produce a caldera-forming eruption on rapid timescales of < 100 ka. This model is unique in that it proposes the possibility of simultaneous melting (wall rock) and crystallization (convecting silicic magma) in the magma chamber. The partial melting region spans from the ambient temperature of the country rock to the magma temperature. It is here that heat is transferred to the crust, and the crust is partially
melted. Further melting takes place in the thermal boundary layer, a zone of heating with a steep temperature gradient. It is here that restitic crystals are resorbed. Local areas of instability create small plumes that may descend from the thermal boundary layer into the hotter convecting magma below. The convecting interior is constantly cooling and crystallizing. Huppert and Sparks (1988) propose that restitic plagioclase crystals from the partially melted crust may become nuclei for new plagioclase growth once in the convecting interior. Fluctuations in temperature within the convecting interior can produce oscillatory zoning.

A more recent model suggests that rhyolites may be derived from crystal mushes in the shallow crust (e.g., Bachmann and Bergantz, 2004, 2008). This model is distinct from the two end-member models, because it addresses the dynamics of a single, cooling magma chamber and does not necessarily require thermal input. The collection of high-silica, aphyric melt at the top of the magma chamber is thought to be the product of settling and compaction of crystals and the ascension of the less dense liquid. The ascension of the silicic liquid is most efficient when a 45 to 65% volume of crystals is present in the magma chamber. Within this window, convection has ceased, and there is high permeability. The collection of melt happens rather quickly, with melt segregation rates of $10^{-1}$ to $10^{-3}$ km$^3$/year. Batholith-sized magma chambers can produce a large volume of rhyolitic magma ($10^2$ to $10^3$ km$^3$) by extracting a small amount ($\leq 10\%$) of the interstitial melt. To produce an eruption, the crystal mush is typically “rejuvenated”, usually by the addition of heat and possibly volatiles from mafic magmas which intrude and pond at the base of the mush. The injection of magma causes overpressurization due to the increase in volume of the contents in the magma chamber. This causes dikes to
rise to the surface, and an eruption occurs. These eruptions may be large-scale or small scale depending upon the amount of rhyolitic melt that has collected at the roof of the chamber. In some cases, the small volume eruptions may be cogenetic, even though several hundred thousand years have lapsed between the eruptions. The system can be reheated multiple times by mafic magma underplating, thus restarting convection and another cycle of crystal formation, compaction, and melt extraction.

The models discussed above have been applied to studies of the three large Quaternary caldera-forming volcanic fields in North America: Yellowstone volcanic field, Wyoming; the Long Valley caldera, California; and the Jemez mountains volcanic field, New Mexico (e.g., Smith, 1979; Hildreth, 1981, 2004; Bachmann and Bergantz, 2008).

1.2 The Valles Caldera

The Valles caldera (~1.23 Ma) is located in the Jemez mountains volcanic field in central northern New Mexico. Two caldera-forming events took place at ~1.61 (Spell et al., 1996) and ~1.23 Ma (Phillips et al., 2007). Post-collapse volcanism began immediately following (~1.23 Ma) the last caldera-forming eruption (Phillips et al., 2007). Post-collapse eruptions have periodically taken place since then (e.g., Bailey et al., 1969; Spell et al., 1993; Spell and Harrison, 1993; Phillips et al., 2007; Gardner et al., 2010). The youngest eruptions of the Valles caldera complex, the East Fork Member (EFM), took place at approximately 55 ka and 40 ka (Goff and Gardner, 2004) after an approximate 460 ka lull in volcanism (Wolff and Gardner, 1995; Wolff et al., 2011).
The Valles caldera complex provides an ideal opportunity for studying the evolution of silicic magma systems because both the caldera-forming ignimbrite (the Bandelier Tuff) and the post-collapse eruptions are well-exposed and preserved. The long quiescence prior to the eruption of the EFM could indicate a change in magmatism and/or perhaps the onset of a new caldera-forming cycle in the Valles caldera complex (Wolff and Gardner, 1995).

1.3 Research Objectives

The following are the main objectives of this study:

(1) *What is the magmatic origin, evolution, and storage condition(s) of the EFM?*

The source of the magma which supplied the EFM eruptions is still unknown. The EFM magma could have been derived from mantle, lower-, and/or upper-crustal reservoirs beneath the Jemez mountains volcanic field. The reservoirs are isotopically distinct from one another (e.g., DePaolo and Wasserburg, 1976; Perry et al., 1987). Pb, Nd, and Sr isotopic compositions of the EFM could elucidate the source(s) of the magma. If two reservoirs were involved, an isotopic mixing model could allow for constraints on the amount derived from each reservoir.

It is possible that the two eruptions of the EFM were derived from different magma bodies. This hypothesis could be tested by the incompatible trace element ratios, which reflect the composition of the magma prior to upper-crustal processes (e.g., crustal assimilation and fractional crystallization). The ratios may be unique from magma batch to magma batch and are usually constrained to a narrow range for a single magma batch.
The evolution and storage conditions of the EFM magma are also unknown. Petrography and crystal compositions can constrain open and/or closed system processes. Rim overgrowths, resorption, and zonation in crystals would strongly indicate that open-system processes were taking place. While residing in the crust, magmas often undergo fractional crystallization and/or crustal assimilation. In most cases, these two processes can be constrained with geochemical data. As stated previously, the radiogenic isotopic compositions of the upper crust are distinct from the other reservoirs. Thus, if upper-crustal assimilation was taking place, the units within the EFM may show a change in isotopic composition over time. If fractional crystallization was taking place, this process could be constrained with whole rock major and trace element data. Major and trace element composition of phenocrysts can define the thermal history and eruption temperature by use of geothermometers.

The eruption age and some aspects of the magmatic history and/or source of the EFM can potentially be constrained by $^{40}$Ar/$^{39}$Ar geochronology. Self et al. (1991) has suggested the potassium-bearing minerals in the EFM may be xenocrystic. If the minerals are indeed xenocrystic, they would not give an eruption age for the EFM but could indicate where they were derived from (e.g., the crust or a primitive magma). If the potassium-bearing minerals or groundmass glass of the EFM contains excess argon, it may indicate a mantle-derived magma was involved in the magmatic history. Mantle-derived minerals and magmas can contain excess argon (Kelley, 2002).
(2) Does the EFM have a discrete magma source when compared to other post-collapse rhyolites in the Valles caldera?

Previous work has shown that the post-collapse eruptions fall into five compositionally different groups (Spell et al., 1993). Furthermore, previous studies on the EFM have demonstrated that the EFM is compositionally and petrologically distinct from the other post-collapse rhyolites (e.g., Self et al., 1988, 1991; Spell et al., 1993; Wolff and Gardner, 1995). The comprehensive petrologic, chemical, and isotopic data collected in this study can provide a detailed test of the hypothesis that the EFM has a discrete magma source, as suggested by previous studies.

(3) Was the event that triggered the eruption injection of basalt into the crust?

Basaltic magmatism has taken place throughout most of the history of the Jemez mountains volcanic field (e.g., Bailey et al., 1969; Gardner and Goff, 1984; Wolff et al., 2005) but has been largely replaced by rhyolitic magmatism over the past 2 Ma. However, basalt may still play a role in generation of magmas within the Jemez mountains volcanic field. Based on a recent study by Wolff and Gardner (1995), the EFM magma may have been generated by the melting of the crust. Their hypothesis is supportive of the Huppert and Sparks (1988) rapid production model (discussed above) as the preferred model for the EFM. The detailed petrographic, isotopic and geochemical data from this study could test that hypothesis.

(4) How can the knowledge of the petrogenesis of the EFM further our understanding of the evolution of large silicic magma systems?

The understanding of post-collapse eruptions is crucial to the understanding of the caldera-forming magma system as a whole. Caldera-forming systems are often multi-
cyclic in terms of magma production and activity; the “post-collapse” eruptions may be “pre-collapse” eruptions for the next caldera-forming cycle. The applicability of the results to the models described above may help to further our knowledge of the probability of future eruptions in the Valles caldera complex.
CHAPTER 2

GEOLOGIC SETTING AND HISTORY

2.1 The Jemez Mountains Volcanic Field

Geographically, the Jemez mountains volcanic field (JMVF) is located in north-central New Mexico, USA. The JMVF overlies the intersection of the Jemez lineament, a northeast-trending Paleoproterozoic suture zone, and the Rio Grande rift, a north-trending zone of Cenozoic extension (Figure 2.01a). The JMVF lies on the western edge of the Española Basin and is built upon Upper Paleozoic sedimentary strata which overlie Proterozoic basement (Smith et al., 1970; Gardner et al., 1986). Regional extension and local faulting have played a vital role in the volcanism in the JMVF. Recent seismic studies suggest that the Jemez lineament is a lithospheric penetrating feature separating a low velocity mantle in the north and a high velocity mantle in the south. The surface separating the two mantle zones dips shallowly southward and is interpreted to be the Southern Yavapai-Mazatzal suture (Shaw and Karlstrom, 1999; Magnani et al., 2004). A low velocity zone which extends from the Moho (approximately 40 km) to the base of the lithosphere (approximately 120 km) could possibly be the source of the magmas for the JMVF (Wolff et al., 2005).

Varying degrees of evolution of mantle-derived magmas and their interaction with crustal rocks has produced basalts, andesites, dacites, and rhyolites in the JMVF. Magmatism began as early as 16.5 Ma (Gardner and Goff, 1984). This early volcanism is recorded by thin flows of alkali basalts interbedded with basin-fill sediments of the Santa Fe Group (Figure 2.02). By 13 Ma, the alkali basalts had been succeeded by the Paliza Canyon andesite, which continued to be erupted until approximately 7 Ma. During this
Figure 2.01. Geologic Map of a) the Valles caldera and b) the East Fork Member. Modified from Wolff et al. (2010).
period of time, there were also minor eruptions of tholeiitic basalt, dacite, the Canovas
Canyon Rhyolite, and the Bearhead Rhyolite (which continued to erupt until
approximately 6.5 Ma), in the southern portion of the JMVF (Gardner et al., 1986; Justet,
1999; Justet and Spell, 2001). In the northern part of the JMVF, the Miocene La Grulla Formation, the Tshicoma Formation andesite, dacite, and rhyodacite (~6.9 to 2.2 Ma), and the Puye Formation (Gardner et al., 1986) are exposed. The Tshicoma, La Grulla, and Paliza Canyon Formations as well as the Bearhead and Canovas Canyon Rhyolites collectively make up the Keres Group (Goff et al., 2011).

2.2 The Valles Caldera

The Valles caldera was first studied in detail by R.A. Bailey, C.S. Ross, and R.L. Smith (Smith et al., 1961; Smith and Bailey, 1966; Smith and Bailey, 1968; Bailey et al., 1969; Smith et al., 1970). Due to their in-depth study, the Valles caldera is now known as the type example of a resurgent caldera. In the Bailey et al. (1969) study, the formal stratigraphy of the Jemez Mountains was defined and the units that make up the Cochiti and Puye Formations as well as the Keres, Polvadera, and Tewa Groups of the Valles caldera were described.

The Valles caldera complex (VCC) is the most prominent feature of the JMVF and is known as the type locality for resurgent dome calderas. The VCC has had two large-scale ignimbrite eruptions followed by caldera collapse and resurgence (discussed below), as first described by Smith and Bailey (1968). The Valles caldera is a semicircular depression with varying depth from 90 to 650 meters. It is approximately 23 km in diameter east-west and approximately 19 km north-south. Within the caldera, a resurgent dome, known as Redondo Peak, rises 3,430 meters above sea level and has over 1,000 meters of local relief. Between the resurgent dome and the caldera rim, post-
collapse rhyolite domes, flows, and associated pyroclastic rocks have erupted along ring fractures.

In general, a caldera complex develops in several stages (Lipman, 2000). In the first stage, the accumulation of magma in a shallow magma chamber causes regional tumescence and the eruption of small domes. The second stage is the initial vent formation, but it is often poorly recorded, if at all, as it is later destroyed by the large-scale eruptions and caldera collapse that occurs. The third stage entails the resurgence and post-collapse magmatism due to renewed magma replenishment in the magma chamber. The fourth stage is that of hydrothermal activity and mineralization, which is present throughout the lifetime of the caldera but is usually the prevailing volcanic activity late in the waning stages of the caldera-forming cycle. This general model can be applied to the VCC.

2.3 The Tewa Group

The history of the Tewa Group, began in the Quaternary begins with the eruptions of the Bandelier Tuff, followed by the eruptive members and units that make up the Valles Rhyolite. The Tewa Group eruptions and subsequent erosion have been the dominant contributor to the topographic and geomorphic landscape that makes up the JMVF.

The Bandelier Tuff is composed of three members and one formation (Figure 2.02). The oldest member of the Bandelier Tuff is the La Cueva Member. It is composed of two discrete rhyolitic ignimbrites which are present in the southwest sector of the Valles caldera and the walls of San Diego Canyon (Self et al., 1986; Gardner et al.,...
The La Cueva Member has been stratigraphically correlated to tephras in the Puye Formation which have yielded $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 1.85 Ma for primary beds (Spell et al., 1990). The La Cueva Member is unconformably overlain by the Otowi Member of the Bandelier Tuff, which erupted at 1.61 Ma (Spell et al., 1996). The post-eruption collapse formed the Toldeo caldera, which was followed by the eruptions of the intracaldera domes and pyroclastic rocks that make up the Cerro Toledo Formation. Within the Cerro Toldeo Formation, there exist four members: the Pueblo Canyon, Alamo Canyon, Virgin Mesa, and Valle Toledo Members. The Valle Toledo Member is composed of several volcanic domes while the other members of the Cerro Toldeo Formation are fluvial deposits (Gardner et al., 2010). The eruption of the Tshirege Member of the Bandelier Tuff occurred at approximately 1.23 Ma (Phillips et al., 2007). The eruption and subsequent collapse partially destroyed the Toldeo caldera and formed the Valles caldera. The Otowi and Tshirege Members constitute approximately 400 to 475 km$^3$ of erupted material each (Cook et al., 2006; Self et al., 2009) of material and culminated in the formation of two nested calderas, the Toldeo and Valles calderas.

Following the Valles caldera collapse, the domes, flows, and pyroclastic beds that make up the Valles Rhyolite were erupted. The members within the Valles Rhyolite are listed here chronologically from oldest to youngest: Deer Canyon, Redondo Creek, Cerro del Medio, Cerros del Abrigo, Cerro Santa Rosa, Cerro San Luis, Cerro Seco, San Antonio Mountain, South Mountain, and the East Fork Members (Figure 2.02). Each member is composed of one or more eruptive units. The earliest post-collapse eruptions were those of the Deer Canyon and Redondo Creek Members at approximately 1.23 Ma (Phillips et al., 2007), almost immediately following caldera collapse. The most recent
eruptions were those of East Fork Member at approximately 55 ka and 40 ka (Goff and Gardner, 2004). Thus, the Valles Rhyolite spans approximately 1.2 Ma.

2.4 The East Fork Member

The most recent eruptive activity to have occurred in the VCC was the eruption of the El Cajete Pyroclastic Beds (ECPB), Banco Bonito Flow (BBF), Battleship Rock Ignimbrite (BRI), and VC-1 Rhyolite, which collectively make up the East Fork Member (Gardner et al., 2010). These units are situated in the southwest sector of the caldera, mainly between the caldera rim and resurgent dome (Figure 2.01b). The east fork of the Jemez River provides good exposures of the section, thus their member name (Gardner et al., 2010). The stratigraphic hierarchy is believed to be as follows: the synchronous or closely-timed eruptions of the ECPB and the BRI along with the possible emplacement of the VC-1 Rhyolite at approximately 55 ka, and the overlying BBF at about 40 ka (Wolff et al., 2011).

The ECPB, the BRI, and the BBF erupted from one or more vents in close proximity to or within the El Cajete crater (Wolff et al., 2011). The three units are petrographically similar to one another, yet petrologically and compositionally distinct from the other preceding post-collapse units (Self et al., 1988, 1991; Spell et al., 1993; Wolff and Gardner, 1995). The East Fork Member has been under careful study within the past few decades as it has been suggested that these eruptions may reflect a new onset of a caldera-forming cycle, if not at least renewed volcanism in the area (Wolff and Gardner, 1995).
CHAPTER 3

PREVIOUS WORK

3.1 The East Fork Member

The units of the East Fork Member (EFM), part of the Tewa Group, were first studied by Bailey et al. (1969). The three units were described individually and given formal stratigraphic order as separate members within the Valles Rhyolite. The former Battleship Rock Member was described as “a sequence of local rhyolitic ash-flow deposits cropping out at the head of the Cañon de San Diego”. The former El Cajete Member was described as “a mantle-bedded air-fall deposit of rhyolitic pumice lapilli and blocks”. The former Banco Bonito Member was described as a “porphyritic obsidian flow that fills the southwestern moat of the Valles caldera”. The initial eruptive stratigraphy of Bailey et al. (1969) is the Battleship Rock Ignimbrite (BRI), El Cajete Pyroclastic Beds (ECPB), and the Banco Bonito Flow (BBF), chronologically from oldest to youngest. The authors noted the three units to be petrographically similar and that the units were likely erupted from the topographical dish-shaped feature named the El Cajete crater based on their distribution. The first geochronologic work was also performed by Bailey et al. (1969), which produced an age of greater than 42 ka by $^{14}$C from charcoal retrieved from a thin ash flow unit which overlies the ECPB. They also presented evidence for an estimated age of younger than 100 ka as supported by the relatively young geomorphic carapace of the BBF.

The East Fork stratigraphy became more complex after cores were retrieved from the Continental Scientific Drilling Program core hole VC-1, which drilled approximately 298 meters into the subsurface below the EFM. The initial description of the core was
made by Goff et al. (1986). In the core, they found at least four welded tuffs, several vitrophyric lavas, and minor pyroclastic fallout beds. They dubbed one of the vitrophyric lavas the “VC-1 Rhyolite”, which was thought to be part of the EFM eruptive sequence due to its petrographic similarities. Self et al. (1988, 1991) attempted to correlate the units to those observed at the surface. A recent study by Wolff et al. (2011) has successfully correlated the stratigraphy observed within the core to the BRI eruptive sequence and BBF. The VC-1 rhyolite remains to be known only in the subsurface. The VC-1 rhyolite was not sampled in this study due to inaccessibility of the materials.

Self et al. (1988) described the petrography of the EFM in detail. They observed that the three units are petrographically similar and that most phenocrysts were not in equilibrium with the rhyolitic melt at the time of eruption. There are crystal aggregates whose origins are likely to be that of melted igneous country rock (Self et al., 1988). Due to the many similarities of the units, the authors grouped them together informally naming them the “El Cajete Series”. This informal name was later retracted by Self et al. (1991) due to possible confusion with the, then current, stratigraphic nomenclature. Self et al. (1988) suggested a $^{40}\text{Ar}/^{39}\text{Ar}$ age range of 170 to 240 ka. They speculated that this age range was too old based on stratigraphic relations in the field and is perhaps due to the dating of xenocrystic material. They also calculated that the BBF lava produced approximately 0.9 km³ DRE (dense rock equivalent) and that the ECPB, BRI, and the VC-1 rhyolite lava totaled to 1.8 km³ DRE.

The EFM is notably different than the other post-collapse eruption products due to its lower silica content as well as the long eruptive hiatus between it and the previous eruption in the Valles caldera, that of the South Mountain Member at 521 ± 4 ka (Spell
and Harrison, 1993). The EFM has been noted to be chemically less evolved than the preceding post-collapse eruptions. The units have lower SiO$_2$ and K$_2$O and incompatible trace elements (Rb, Cs, HREE, Pb, Th) with higher amounts of Ti, Fe, Ca, Zr, Sr, Ba, and Eu (Spell et al., 1993).

The possibility of a common petrogenesis for the three units was later noted by Wolff and Gardner (1995), who named the units the “Southwestern Moat Rhyolites” due to their location in the southwestern sector of the caldera between the resurgent dome and caldera rim. The authors observed many of the same textures as Self et al. (1988) and deduced that the disequilibrium textures and restitic crystals are indicative of magma mixing between a basaltic andesite and a rhyolitic magma.

Several recent geochronologic studies have attempted to obtain an age for the East Fork eruptive sequence. Toyoda et al. (1995) report an age of 59 ± 6 ka for the BRI and 53 ± 6 ka for the ECPB by use of electron spin resonance on quartz retrieved from the two units. Reneau et al. (1996) present an age range of 47 to 65 ka for the ECPB and BRI using thermoluminescence and high-precision $^{14}$C. A study performed by Phillips et al. (1998) yielded a $^{21}$Ne exposure age of 44 ± 11 ka of the soil overlying the ECPB, thus providing an lower age limit for the EFM. Based on this recent geochronologic work, the reported ages range from approximately 60 to 40 ka (Gardner et al., 2010). More specifically, the best age estimate for the ECPB and BRI units is 55 ± 6 ka and 40 ± 5 ka for the BBF lava due to the repetition of the age within the obtained ranges by the various dating techniques (Goff and Gardner, 2004; Wolff et al., 2011). A more recent attempt to date the baked sediments covered by the BBF using the optically stimulated luminescence method was inconclusive (Lepper and Goff, 2007).
The most recent revision to the stratigraphy, not only of the EFM but also of the entire stratigraphy of the VCC, was made by Gardner et al. (2010). The revision proposes that the previous informal names and former member names (El Cajete, Banco Bonito, and Battleship Rock Members) be abandoned in favor of creating a single member name for the three units: the EFM. Detailed physical descriptions of the stratigraphic units that compose the EFM can be found in Wolff et al. (2010).

In conclusion, while some aspects of the EFM stratigraphy remain imperfectly known, the EFM was and is still recognized to be the youngest eruption in the Valles caldera. In fact, the eruption of the ECPB eruption was the last plinian style eruption to have occurred in the mid-continent United States (Reneau et al., 1996).
CHAPTER 4

SAMPLING STRATEGY AND ANALYTICAL METHODS

4.1 Sample Collection

Field work and sample collection for this study took place during late July and early August 2009. The eruptive units of the Banco Bonito Flow (BBF), El Cajete Pyroclastic Beds (ECPB), and Battleship Rock Ignimbrite (BRI) were sampled based on the mapping of Smith et al. (1970) and Wolff (personal communication, 2009). In total, forty samples were collected from road-cuts along NM Highway 4 and natural outcrops (Figures 4.01, 4.02, 4.03). Of those forty collected, twenty-seven samples were selected for thin section preparation and petrographic analysis as well as possible isotopic, major and trace element analysis. Hand samples were first examined macroscopically to exclude any altered samples as well as to make preliminary observations of the crystal assemblages. Since some of the samples were collected close to one another, the most representative and unaltered sample from each area was chosen for thin section preparation. All other samples were set aside for possible future use.

The samples were selected for isotope and geochemical analysis by the criteria that they appeared to be unaltered, were of the appropriate size, and were representative of the unit they were sampled from. The samples showed no flow banding. After petrographic inspection in thin section, smaller subsets of these samples were selected for X-ray fluorescence (XRF) and inductively coupled plasma mass spectrometry (ICPMS), thermal ionization mass spectrometry (TIMS), $^{40}$Ar/$^{39}$Ar geochronology, and electron probe microanalysis (EPMA).
Figure 4.01. Sample locations for the El Cajete Pyroclastic Beds. The small elliptical feature is the Las Conchas Quarry. New Mexico Highway 4 cuts across the figure. The image is modified from Google Earth.

Criteria for selecting the samples to be analyzed by XRF and ICPMS was based on their location in the stratigraphy of the ECPB (one sample from the base of the unit, two from the middle, and one sample from the top), the proximity to the vent and distal locations on the bifurcated flow lobes of the BBF, and a single large, non-welded pumice clast near the base of the BRI (Appendix A). It should be noted the sample name scheme is based upon the unit, stratigraphy within the unit (only applicable to the ECPB), and sequential collection number. For example, ECH01 denotes the first sample collected
Figure 4.02. Sample locations for the Battleship Rock Ignimbrite. The road in the figure is New Mexico Highway 4. The image is modified from Google Earth.

from stratigraphic level H in the ECPB. To prevent confusion, the alphanumeric nomenclature of Wolff et al. (2010) has been retained and employed in the naming scheme used in this thesis. Sample selections for each analytical method and UTM coordinates are listed in Appendix A.

4.2 Sample Preparation

Selected hand samples were cut using a rock saw to standard thin section billet dimensions (~22 x 44 x 15 mm). The pumice samples that were too fragile for typical
Figure 4.03. Samples locations for the Banco Bonito Flow. The elliptical feature is the El Cajete Crater. New Mexico Highway 4 cuts across the figure. The image is modified from Google Earth.

Sample preparation were first impregnated with epoxy before they were cut into billets. Twenty-seven billets were sent to Quality Thin Sections in Tucson, Arizona to be cut into thin sections. The pieces of the hand sample that each billet was cut from were set aside for possible future use.

4.3 Petrography

Prior to microscope work, the hand samples were examined macroscopically to describe their general texture, crystal size, and to identify visible minerals. In thin
section, textures as well as crystal assemblages and abundances were noted in order to identify major, minor, and accessory minerals. This strategy also provided a detailed petrographic description for each sample. If present, alteration, glass inclusions, reaction rims, zonation, and crystal intergrowths were noted in petrographic descriptions. In addition to the general petrography, point counts were conducted on a sample representative of each unit. Approximately 500 points per thin section were counted to calculate proportions of crystals and groundmass material.

4.4 Electron Probe Microanalysis

Thirteen thin sections were selected for analysis using EPMA. The thin sections were selected based on the criteria of the presence and abundance of the desired crystal assemblages they contained. The purpose of analyzing the samples using EPMA was to determine the composition of, and chemical variations within, crystals and glass inclusions, if present. Targeted crystals were circled with a permanent marker on the reverse (glass) side of the thin section. The targeted crystals were feldspar, amphibole, biotite, quartz, and opaque oxides. Prepared thin sections were polished by hand using a sequence of 9, 6, 3, and 1 µm diamond polishing spray on a polishing pad placed on a pane of glass. The samples were rinsed with water and a new polishing pad was used at each step to minimize scratching of the polished surface. When polishing was complete, the samples were rinsed with water a final time to remove any polishing debris.

To ensure that the electron charge deposited on the sample during analysis would be uniformly conducted to ground, the thin sections were carbon coated. The thin sections were placed face upward in a bell jar, which was then evacuated to less than $10^{-5}$
torr. Current passing through the carbon electrodes caused the carbon to vaporize, and a thin carbon film was deposited on the thin section surface.

All EPMA work was performed at the Electron Microanalysis and Imaging Laboratory at UNLV using a JXA-8900 SuperProbe. The selected samples were viewed with backscatter images to detect variations in the mean atomic number in order to identify zoning and other chemical variations within the crystal. The backscatter image is produced when the beam of electrons produced by the microprobe encounter the sample and the high energy backscattered electrons rebound off the surface. The quantity of backscatter electrons is proportional to the mean atomic number of the sample under the electron beam. The higher the mean atomic number, the brighter the crystal appears in the backscatter image.

Major element data were collected from selected points and transects on targeted crystals and glass inclusions. Points on the crystals, such as obvious overgrowth rims or zonation, were first analyzed to ensure the data collected would be acceptable before performing line transects. Point data, backscatter images, photomicrographs taken on a petrographic microscope prior to the EPMA analyses, and observations made throughout the process were used to pick crystals suitable for EPMA analyses.

4.4.1 Feldspar

A majority of the feldspars in all three units are characterized by resorbed cores overgrown by either less resorbed or unresorbed rims. EPMA line transects were performed across feldspar crystals with obvious rims. Most transects were performed either rim-to-core or rim-to-rim, if circumstances allowed. Since most of the feldspars were resorbed, the ability to do rim-to-rim transects was hindered and limited the
analyses that could be done. For those samples with a portion of the feldspar crystal missing (due to the cut of the thin section or to extremely heavy resorption), a line transect was performed from the rim to the interior of the crystal. The feldspar crystals analyzed are found in samples ECH01, BR05, BB01, BB04, BB15, BB17, and BB24. The brown glass inclusions within two feldspars in BR05 were also analyzed for major element chemistry.

### 4.4.2 Amphibole

Point analyses and line transects were performed on amphibole in samples BB19 and ECH01. BB19 showed a high abundance of euhedral amphibole crystals in thin section. Larger crystals were selected for rim-to-rim transects while individual point analyses were conducted on small crystals. If the smaller crystals were large enough, several points were taken on the rims and the core. No amphibole was chosen from the BRI due to the lack of analyzable amphibole (i.e. the crystals were either too heavily resorbed or small).

### 4.4.3 Biotite

The biotite crystals in samples BB01, BB17, and BB24 were analyzed by rim-to-rim transects. These samples were chosen due to the high abundance of biotite in the samples. No biotite samples were analyzed from the ECPB and the BRI due to the lack of analyzable biotite.

### 4.4.4 Quartz

Core-to-rim transects were performed across the subhedral quartz crystals found in BB04 and BB24 in order to collect trace element titanium data. The samples were
chosen due to the high abundance of quartz crystals. No other quartz crystals were analyzed as most were heavily resorbed or were fragments.

4.4.5 Oxides

The oxides selected for EPMA were usually present as euhedral opaques in crystal aggregates. Since the oxide crystals were so small, point analyses were performed on individual crystals. The oxides analyzed were from samples BB04 and BB19.

4.5 Geothermometry by EPMA

Two geothermometers, the titanium-in-quartz (Wark and Watson, 2006) and Fe-Ti geothermometer (Buddington and Lindsley, 1964; Ghiorso and Evans, 2008) were applied in this study to not only determine the magma temperature prior to eruption but also to elucidate the variability in magma temperature of the EFM magma prior to eruption. The titanium-in-quartz geothermometer was the primary geothermometer in the study. The Fe-Ti oxide geothermometer was used to confirm the validity of the temperatures obtained by the titanium-in-quartz geothermometer. The titanium-in-quartz geothermometer can record temperature(s) of the magma by the proportion of Ti substituted into the Si tetrahedral site in the quartz crystal structure. The Fe-Ti two oxide geothermometer calculates temperature based on the exchange of $\text{Fe}^{2+} \text{Ti} \leftrightarrow (\text{Fe}^{3+})_2$ or $\text{Fe}^{2+} \leftrightarrow \text{Mg}$ between the two oxides.

4.6 XRF and ICPMS Analyses

Fresh rock chips of the samples selected for ICPMS and XRF were taken from the remnants of the hand sample from which a billet was cut, if available. These selections
were made to ensure that the chips were not only free from weathered rinds and other such defects, but also that the whole rock chemistry would closely represent what can be seen in the thin section. If there were no remnants of billet cuttings left over, rock chips were taken from hand samples. Approximately ten grams of each of the twelve selected samples were sent to the Washington State University GeoAnalytical lab and were further processed and analyzed to measure major, trace, and rare earth elements following the methods described by Knaack et al. (1994) and Johnson et al. (1999) for ICPMS and XRF, respectively.

Loss on ignition (LOI) calculations were also performed at Washington State University as a part of the XRF and ICPMS analytical process. LOI calculations measure the amount of water and volatiles present in each sample and allow for estimation as to the reliability of the data due to the mobility of certain elements.

4.7 \(^{40}\text{Ar}/^{39}\text{Ar}\) Geochronology

Two samples (BB17 and BB19) were chosen for geochronologic work. BB17 was selected due to the relatively pristine state of the groundmass glass as compared to the other glassy rhyolite samples. BB19 was chosen due to its high abundance of large, euhedral biotite crystals seen in thin section. For each selection, a hand sample was cleaned of weathered rinds and crushed to approximate 3 cm diameter chips using a jaw crusher. The rock chips were further crushed using a rotary disk mill. The crushed material was then sieved on a stack of 710, 600, 540, 425, and 250 µm mesh sieves. Each fraction separated was stored in individual containers, and later the 600 and 540 µm size fractions were picked by hand using a binocular microscope to select inclusion-free,
euhedral biotite and clean glass for BB19 and BB17, respectively. Approximately 250
mg of each sample were sent to the United States Geological Survey TRIGA Reactor in
Denver, CO for irradiation. The fluence monitor used was a Fish Canyon Tuff sanidine,
FC-2, with an age of 28.02 ± 0.16 Ma (Renne et al., 1998). The samples were placed in a
Cd-lined in-core irradiation tube. The tube was placed vertically in the center of the rod
and irradiated for one hour. The returned irradiated samples were separated into two
individual batches of glass and two individual batches of biotite. The batches were
analyzed at the Nevada Isotope Geochronology Lab at the University of Nevada, Las
Vegas.

Each aliquot was analyzed separately, but there were common analytical
procedures among all the samples. Each aliquot was step-heated in the furnace using a
heating schedule based on the material properties. The sample was held at a specific
temperature for 12 minutes while diffusion allowed for the argon gas once trapped in the
biotite crystal lattice or glass to be released into the extraction line which purified the
gases using SAES getters and finally into the mass spectrometer. The final temperature
each sample reached was 1400°C, at which any of the remaining argon gas not released
during the previous heating steps would be released. This final step is called the fusion
step. The analyses had the following correction factors: 4 amu discrimination of 1.0549
± 0.55%, $^{40}\text{Ar}/^{39}\text{K}$ of 0.0002 ± 150.0%, $^{36}\text{Ar}/^{37}\text{Ca}$ of 0.0003573 ± 13.29%, and $^{39}\text{Ar}/^{37}\text{Ca}$ of 0.000641 ± 20.61%.

BB17 biotite aliquot #1 weighed 110.0 mg. It was heated in 14 steps at 650, 725,
790, 850, 905, 960, 1015, 1050, 1085, 1115, 1145, 1180, 1220, and 1400 °C. BB17
biotite aliquot #2 weighed 105.5 mg. Aliquot #2 was heated in 12 steps at 790, 860, 920,
960, 1010, 1040, 1070, 1100, 1130, 1160, 1200, and 1400 °C. The reason for this change in step heating temperatures for aliquot #2 was because the first aliquot (BB17 aliquot #1) released too much of the argon gas in too few steps (steps 7 through 10). BB19 glass #1 weighed 98.10 mg. BB19 glass aliquot #2 weighed 106.6 mg. Both of the glass batches were step heated separately at 675, 750, 800, 820, 840, 865, 890, 950, 1000, 1050, 1100, 1150, and 1400 °C.

In general, there are two preferred methods to define an age using \(^{40}\text{Ar}/^{39}\text{Ar}\) geochronology: a plateau age and an isochron age. A plateau age is calculated when three or more consecutive steps are within 2-sigma uncertainty (95% confidence level) of each other and represent greater than 50% of the total gas released from the sample. “Ideal” behavior occurs when the age spectrum forms a single plateau in which each step is indistinguishable from the others. This behavior is indicative of a simple geologic history (i.e. quick cooling with no reheating or alteration). A discordant age spectrum would indicate a more complex history. Samples that return “saddle-shaped” age spectra have often been associated with excess \(^{40}\text{Ar}\) (McDougall and Harrison, 1999).

An isochron age is preferred to a plateau age as it gives more information (e.g., the initial, non-radiogenic argon present in the sample). This initial argon composition should be equal to the current value for atmospheric argon (295.5). For reasons of statistical reliability, four or more points are desirable to produce an isochron.

The total gas age is the mean age of all the individual heating steps and is calculated by weighing the individual steps by the fraction of \(^{39}\text{Ar}\) released. This age is analogous to that of the conventional K/Ar age.
4.8 Thermal Ionization Mass Spectrometry

A representative sample from each unit was selected for TIMS Pb, Nd and Sr isotopic analysis. Clean rock chips were selected from the pieces left over from cutting the billets. The samples were powdered by a tungsten-carbide shatter-box (BR01 and BB01) or by hand using a corundum mortar and pestle (ECH01). The individual powders were sealed in their own vials and then shipped to the University of Colorado, Boulder for analytical work in their TIMS lab.

The three samples were analyzed by a Finnigan-MAT 261 with 6 collectors. During the study period, the total procedural blanks averaged approximately 1 ng for Sr and Pb, and 100 pg for Nd. Fifty-five measurements of SRM-987 yielded a mean of $^{87}\text{Sr}/^{86}\text{Sr} = 0.710292 \pm 1$. Reported errors are 2-sigma of mean. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were analyzed using four-collector static mode measurements. Thirty-two measurements of the La Jolla Nd standard ran during the study period yielded a mean of $^{143}\text{Nd}/^{144}\text{Nd} = 0.511840 \pm 1$. $^{143}\text{Nd}/^{144}\text{Nd}$ analyses were analyzed using dynamic mode, three-collector measurements. Measured $^{143}\text{Nd}/^{144}\text{Nd}$ normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. $\varepsilon_{\text{Nd}}$ values were calculated using a present-day $^{143}\text{Nd}/^{144}\text{Nd}$ (CHUR) = 0.512638. The Pb analyses were four-collector static mode measurements. Corrections for each Pb ratio was made by 16 measurements of standard SRM-981 which resulted in $^{208}\text{Pb}/^{204}\text{Pb}$ values of 36.56 ± 0.03 and $^{207}\text{Pb}/^{204}\text{Pb}$ values of 15.449 ± 0.008. The calculated $^{206}\text{Pb}/^{204}\text{Pb}$ values were 17.65, 17.66, and 17.64 for the BR01, ECH01, and BB01, respectively.
Crystal assemblages are similar for all three units in the East Fork Member (EFM) and are generally as follows (in decreasing abundance): plagioclase, biotite, clinopyroxene, orthopyroxene, amphibole, sanidine, quartz, and opaque oxides as well as minor amounts of apatite and zircon. The apatite and zircon crystals are present both as inclusions within biotite or as individual crystals.

Two distinct crystal populations are present in all three units: the large (≥ 1 mm), resorbed or subhedral crystals representing one population and the small (≤ 0.5 mm), euhedral crystals representing the other population. Crystals that are present in these two populations include: plagioclase, biotite, amphibole, clinopyroxene, orthopyroxene, and sanidine.

Plagioclase is the most abundant mineral phase present in all three units. A majority of the large, resorbed plagioclase grains have rim overgrowths. The rim overgrowths are of the same mineral phase as the interior in most cases. Many of the resorbed plagioclase crystals in the Banco Bonito Flow (BBF) and the Battleship Rock Ignimbrite (BRI) contain brown glass inclusions. The small plagioclase crystals are typically euhedral and have no rim overgrowths. Sanidine exhibits much of the same morphology as that of plagioclase; the large crystals are often resorbed and have rim overgrowths. The small euhedral sanidine crystals typically have no overgrowths and display concentric oscillatory zonation. Biotite is the second-most abundant mineral in the BBF, whereas clinopyroxene is the second-most abundant phase in both the El Cajete Pyroclastic Beds (ECPB) and the BRI. Quartz with undulose extinction is observable in
all three units. No euhedral quartz was observed in any of the units. Of the three units, quartz was most prevalent in the BBF, appearing in nine thin sections.

5.1 Petrography of the El Cajete Pyroclastic Beds

The ECPB are composed of pumice fragments (lapilli) and ash of either fallout or pyroclastic surge deposits. The pumice can be classified as sparsely porphyritic (Table 5.01). In general, the pumice becomes gradually more porphyritic upsection. The dominant minerals present include: plagioclase, biotite, clinopyroxene, orthopyroxene, sanidine, quartz, oxides, and amphibole. Accessory minerals include oxides, apatite, zircon, and olivine. Most of the zircons observed were inclusions within biotite with pleochroic halos around the inclusions. Apatite was also observed as inclusions in biotite.

Samples ECA01, ECA02, ECB01, ECC01, ECD02, ECF01, and ECI01 exhibit clay alteration of the glassy groundmass. The samples that exhibit the highest degree of alteration where most of the glass has decomposed to clay are ECA01, ECC01, and ECF01. Small translucent red blebs, which are likely magnetite crystals that have been oxidized, are present in samples ECA01, ECA02, ECB01, ECC01, and ECD02.

All of the plagioclase and sanidine crystals exhibit zonation. The zonation within both the large and small crystal populations is usually oscillatory and observable in the individual crystals and rim overgrowths, if present. Polysynthetic and simple twinning is common for the plagioclase crystals. The average length of the small plagioclase crystals is approximately 0.3 mm. The average length of the large crystals is approximately 1.75 mm. The larger plagioclase crystals typically have moderately to heavily resorbed cores.
with rim overgrowths displaying slight to no resorption. The large plagioclase crystals are sometimes so heavily resorbed that a skeletal texture is formed. This crystal morphology is observed in ECB01. The small crystals, however, are euhedral and exhibit little to no resorption. They do not have overgrowths and are usually found as individual crystals. The small crystals often display concentric zonation.
The observations of sanidine are similar to those of plagioclase. Large sanidine crystals are approximately 1 mm in length, are heavily resorbed, and some have rim overgrowths. The overgrowths are of the same phase as that of the interior (i.e. no rapikivi texture was observed). The smaller sanidine crystals are often euhedral, zoned, show little to no resorption, and have no rim overgrowths. Both of the large and small sanidine crystal populations occur as individual crystals or in crystal aggregates.

The largest biotite crystals in the ECPB are approximately 2 mm in diameter. They are often subhedral due to resorption. The small biotite crystals are typically euhedral to slightly subhedral. Very small (~0.125 mm) biotite crystals are slightly more abundant than the large crystals. The large biotites are usually present in crystal aggregates but are sometimes observed as individual crystals.

Crystal aggregates were observed in ECA01, ECA02, ECB01, ECD02, ECE01, ECG02, ECH01, and ECI01. These aggregates were usually composed of two or more of the following minerals: plagioclase, clinopyroxene, biotite, sanidine, amphibole, and oxides. The crystals within the aggregates are commonly resorbed.

Notably, a large olivine (approximately 2 mm in diameter) was observed with a reaction halo of clinopyroxene in ECF01. The olivine crystal was heavily resorbed. Smaller sized (0.25 to 0.5 mm) euhedral crystals of clinopyroxene are present both as a rim around the olivine as well as individual crystals in the pumice matrix. Another notable feature is three plagioclase crystals, which contain small brown glass inclusions in ECH01. This feature is otherwise not observed in the ECPB.
5.2 Petrography of the Battleship Rock Ignimbrite

The BRI is a rhyolitic tuff. As is common in ignimbrites, the unit becomes more densely welded in the middle of the unit. In general, the pumice in the BRI can be classified as sparsely porphyritic (Table 5.01). The minerals present include: plagioclase, biotite, amphibole, oxides, quartz, orthopyroxene, and clinopyroxene. Zircon is an accessory mineral and is present as individual crystals or inclusions in biotite.

In general, the pumiceous glass of the BRI is light brown. Individual pumice fragments are present within BR03 and BR05. The pumice fragments are composed of either brown or clear glass and are most prominent in BR03 (Figure 5.01). Lithic fragments are also present in BR03 and are volcanic rocks likely of dacitic or rhyolitic composition. Many of the plagioclase crystals are heavily resorbed and contain brown glass inclusions (Figure 5.02). The plagioclase crystals range in size from >2 mm to 0.05 mm. There is also a small (0.4 mm) microcline crystal that exhibits tartan twinning, undulose extinction, and contains brown glass inclusions in BR03. Some of the subhedral quartz exhibits undulose extinction as well. Resorbed biotite crystals with reaction rims of amphibole are present in BR05. The crystals and rims are approximately 0.2 mm in diameter. There are also zircons (approximately 0.0625 mm in length) in embayed biotite crystals. Small (approximately 0.25 mm to 0.65 mm) euhedral orthopyroxene crystals are present as inclusions within larger euhedral plagioclase and as individual crystals. There are a few fractured quartz fragments.
Figure 5.01. Glass inclusions in plagioclase in BR03. Brown glass inclusions in a heavily resorbed plagioclase with rim overgrowths photographed in PPL (upper) and XPL (lower). Note the brown groundmass glass, anhedral quartz crystals, and clear glass pumice fragments.
Within the unwelded pumice collected at the base of the BRI (sample BR01), many of the large crystals appear to be fragmented while the small crystals remain intact as individual crystals. The plagioclase crystals are often small (< 1 mm), heavily resorbed, and contain brown glass inclusions. Polysynthetic twinning is common. Small (approximately 0.25 mm in length) subhedral sanidine crystals have simple twinning and oscillatory zoning. The moderate to large (approximately 0.5 mm to 1 mm in length) orthopyroxene crystals are often resorbed. An extremely resorbed clinopyroxene with amphibole overgrowth is present. The resorbed crystal with the reaction rim is 2.2 mm in diameter. Individual zircon crystals are also present in this sample.

5.3 Petrography of the Banco Bonito Flow

The BBF is a porphyritic rhyolite ranging from pumiceous to glassy in texture (Table 5.01). In general, the groundmass glass is clear and unaltered. Samples BB04, BB05, BB10, BB12, BB13, BB15, BB17, BB18, and BB22 are pumiceous. Minor clay alteration was observed along the vesicles in BB22.

In general, the crystal assemblages in the BBF samples include: plagioclase, biotite, amphibole, sanidine, orthopyroxene, clinopyroxene, quartz, and oxides. The accessory minerals zircon and apatite are usually present as inclusions within euhedral and subhedral biotite. There is a nonsystematic, slight variation in crystal assemblages and abundances with location in the flow.

Most small (approximately 0.25 mm in length) crystals are euhedral and sometimes flow aligned in the glassy samples. The larger (> 0.5 mm) crystals are often heavily resorbed, some to the point of exhibiting skeletal morphology. The skeletal
crystals are usually feldspars, although skeletal orthopyroxene was observed (BB15). A few of the large quartz and biotite crystals are embayed.

Some notable crystal textures within the BBF include: quartz with undulose extinction in BB17 and resorbed biotite with amphibole reaction rims in BB01 and BB15 (Figure 5.02). Many quite large (≥ 2 mm in length), euhedral biotite crystals are present in BB15 and BB17. Sample BB12 contains the most heavily resorbed crystals, which include plagioclase, sanidine, biotite, quartz, and oxides. In that particular sample, many of the crystals of plagioclase and sanidine exhibit skeletal morphology, and a large biotite is embayed.

Crystal aggregates are common in the BBF. There are two populations of the crystal aggregates: those that are euhedral and others that are resorbed and subhedral. A range of small to large (0.25 mm to ~1.5 mm) aggregates of euhedral crystals are present in sample BB01, BB04, BB12, and BB17. However, the more common aggregates are composed of large (≥ 2 mm), subhedral and resorbed crystals (samples BB05, BB08, BB10, BB12, BB15, BB19, BB24, and BB26) (Figure 5.03). At times, these two types of crystal aggregates occur within the same sample (BB12). The aggregates are usually made up of two or more of the following minerals: plagioclase, sanidine, biotite, oxides, orthopyroxene, amphibole, and/or very minor amounts of red translucent blebs within the resorbed plagioclase.

The plagioclase and sanidine crystals are usually zoned. Simple twinning is common in all the crystals regardless of size. Very small (0.125 mm) individual crystals of amphibole, oxides, and biotite are seen within large resorbed plagioclase in samples BB04, BB10, BB15, and BB17. Sample BB13 displays the most variety of textures of
Figure 5.02 Amphibole reaction rim on biotite in BB15. Amphibole reaction rim around heavily resorbed biotite photographed in PPL (upper) and XPL (lower).
Figure 5.03. Crystal aggregate in BB10. A crystal aggregate composed of plagioclase and pyroxene crystals. Photographed in PPL (upper) and XPL (lower). Note the resorption of and rim overgrowths on the plagioclase crystals.
plagioclase and sanidine seen in the BBF. In the thin section, a large (2.1 mm by 0.5 mm) euhedral sanidine crystal with oscillatory zoning is observable, which is otherwise not observed in the BBF since all the large crystals are resorbed and subhedral. In the same sample, there is a large plagioclase crystal that has a resorbed core with an unresorbed rim overgrowth, a large (1 mm) resorbed sanidine crystal, and a skeletal plagioclase crystal. There are several small (< 0.5 mm) euhedral plagioclase crystals, but they are not as common as in some other samples. One large (approximately 2 mm in length) subhedral, highly resorbed microcline crystal exhibits undulose extinction. Such a spectrum of feldspar textures is not observable in any other sample within the BBF.

Some resorbed sanidine and plagioclase crystals contain brown glass inclusions, although this feature is less common than what is observable in the BRI. Sample BB01 contains a high abundance of resorbed plagioclase with brown glass inclusions (Figure 5.04). Two large glass inclusions measuring approximately 0.25 mm each were observed in one plagioclase crystal from that sample.

Quartz is present in the following samples: BB04, BB05, BB10, BB12, BB13, BB15, BB17, BB18, and BB24. The quartz crystals in the BBF are generally subhedral with occasional anhedral crystals. They occur as individual crystals and range in size from approximately 0.25 to 2.25 mm.

Euhedral amphibole is more prevalent in the BBF than in the ECPB and BRI. Resorption is exhibited by only a few crystals. Amphibole is present in samples BB04, BB08, BB10, BB12, BB15, BB17, BB19, BB24, and BB26. The small (0.25 mm) crystals are often flow aligned. The largest amphibole crystals (up to 2 mm in length) can be viewed in sample BB19. Several small (0.5 mm) euhedral amphibole are present
Figure 5.04. Brown glass inclusions in two resorbed plagioclase crystals in BB01. Both crystals were photographed in PPL.

as inclusions within a large (> 4 mm) resorbed sanidine that has an unresorbed overgrowth rim in BB17. Amphibole is present in crystal aggregates in BB15 and BB17.
CHAPTER 6

RESULTS

6.1 Electron Probe Microanalysis

Data tables for all samples analyzed by EPMA can be found in Appendices B through F. Note that all molar percentages discussed below are averages and uncertainties are 1-sigma unless otherwise denoted. All major elements are expressed as weight percent.

6.1.1 Plagioclase Feldspar

As described in Chapter 5, there are two general populations of crystals in the East Fork Member (EFM). In thin section, oscillatory zoning is noted in both the large and small plagioclase populations. However, rim-to-rim and rim-to-core EPMA transects were only performed on the large plagioclase crystals, particularly those with rim overgrowths. EPMA data tables for the analyzed feldspar can be found in Appendix B.

Three different types of zonation are observed in the plagioclase crystals from the BBF and the ECPB samples: normal, reverse, and patchy. Examples of the three types of zonation can be seen in Figure 6.01. Plagioclase crystals with reverse zonation are characterized by sodic cores and calcic rims and are found in samples BB01 and BB15. Plagioclase with normal zonation exhibit calcic cores with sodic rims and can be seen in samples BB17 and BB24. Plagioclase exhibiting patchy zonation can be found in ECH01. No other feldspars were analyzed from the ECPB or the BRI as almost all were heavily resorbed or too small.
Figure 6.01. EPMA transects across plagioclase crystals from the El Cajete Pyroclastic Beds and Banco Bonito Flow. Analyses start at the rim of the crystal. Percent albite, anorthite, and orthoclase denoted by symbols.

Reverse zonation is displayed by samples BB01 Plag1 and BB15 Plag1 (Figures 6.01 and 6.02). The core composition of BB01 Plag1 is oligoclase (An16 to An19) and averages An17. The rim composition ranges from oligoclase (An25 to An28) and averages An26. The core composition of BB15 Plag1 is oligoclase to andesine (An17 to An33) and averages An22. The rim composition is oligoclase (An28 to An29) and averages An28.
Figure 6.02. Ternary diagrams of plagioclase with reverse zonation from the BBF. Note: or = orthoclase, ab = albite, an = anorthite. △ = core    ○ = rim
Normal zonation is shown in samples BB17 Plag2 and BB24 Plag2 (Figures 6.01 and 6.03). The core of BB17 Plag2 is andesine (An$_{31}$ to An$_{40}$) and is overgrown by a rim of oligoclase to andesine (An$_{24}$ to An$_{33}$). The core and rim averages are An$_{35}$ and An$_{28}$, respectively. The core of BB24 Plag2 is andesine (An$_{48}$ to An$_{36}$). The rim overgrowths are andesine (An$_{33}$ to An$_{39}$). The core and rim averages are An$_{42}$ and An$_{36}$, respectively.

A plagioclase crystal in sample BB01 has an obvious rim of higher anorthite content (An$_{26}$) compared to the core (An$_{17}$). A diffusion profile for Mg was attempted to be defined by running several transects across the core-rim boundary. The profile was an effort to elucidate whether Mg was diffusing across the overgrowth boundary in order to estimate when the new growth had occurred. A point analysis was performed every 6 µm. However, the Mg concentrations were below detection limits, and thus no useful data were obtained.

### 6.1.2 Amphibole

To determine if compositional zonation was present in amphibole, small to large euhedral crystals were analyzed from the BBF and the ECPB. No amphibole crystals were analyzed from the BRI because they were heavily resorbed. EPMA data tables for the analyzed amphibole and biotite can be found in Appendix C. Amphibole crystals were analyzed in samples BB01, BB08, BB17, BB19, and ECH01. All analyzed amphibole were individual crystals with the exception of a crystal aggregate in BB17. Point analyses were taken on the core and rims of the small crystals, and line transects were performed across the large crystals. No significant change in MgO or TiO$_2$ is evident in the points analyzed at the rim and core of the smaller amphibole crystals from the BBF samples. The large crystals showed little to no compositional zonation in rim to
Figure 6.03. Ternary diagrams of plagioclase with normal zonation from the BBF. 
Note: or = orthoclase, ab = albite, an = anorthite. △ = core  ⬤ = rim
rim transects. Examples are shown in Figure 6.04. The amphibole in sample ECH01 are slightly more MgO and TiO$_2$ rich the other analyzed amphibole. In that sample, the average MgO and TiO$_2$ is $16.44 \pm 0.72$ and $2.05 \pm 0.40$, respectively. The average MgO and TiO$_2$ for the BBF amphibole is $15.18 \pm 0.83$ and $1.57 \pm 0.56$, respectively. The highest MgO and TiO$_2$ content is found in ECH01 Amphibole2 and Amphibole3, which have an average of $16.48 \pm 0.78$ for MgO and $2.18 \pm 0.27$ for TiO$_2$. The average MgO and TiO$_2$ for amphiboles from both units is $15.44 \pm 0.95$ and $1.67 \pm 0.56$, respectively. The average Mg number for all of the analyzed amphibole is $68.01 \pm 2.84$.

![EPMA transects across biotite and amphibole from the El Cajete Pyroclastic Beds and the Banco Bonito Flow.](image)

**Figure 6.04.** EPMA transects across biotite and amphibole from the El Cajete Pyroclastic Beds and the Banco Bonito Flow.

### 6.1.3 Biotite

Biotite was analyzed to determine if any compositional zonation was present. Appendix C contains EPMA data for biotite and amphibole. Large, euhedral biotite
crystals were selected from the BBF. No crystals were chosen from the ECPB or BRI because the samples were either heavily resorbed or too small. The selected BBF biotite crystals were either independent crystals (sample BB01) or parts of crystal aggregates (samples BB17 and BB24). The rim-to-rim transects revealed no significant zonation. Examples of TiO$_2$ and MgO variation are shown in Figure 6.04. The average Mg number is 66.52 ± 1.46, and the average MgO is 15.61 ± 0.74. The highest MgO is seen in BB01 biotite1. In the same crystal, the highest TiO$_2$ is also observed, which averages 4.74 ± 0.12. The other BB01 biotite samples average 1.41 ± 0.22 TiO$_2$.

6.1.4 Glass Inclusions

Single point analyses were performed on brown glass inclusions in two plagioclase crystals in sample BR05 from the BRI. To determine if a trend in composition was evident, inclusions were selected and analyzed from the core to the rim of the crystal. The analyses yielded no compositional variance with position within the feldspar crystal. Glass inclusion data can be found in Appendix D.

Using the classification of Le Bas et al. (1986), the glass inclusion data reveals two distinct populations (Figure 6.05). One population plots in the trachy-andesite field (average 58.62 SiO$_2$) and the other in the rhyolite field (average 72.55 SiO$_2$). There is a single data point that plots at 68.70 SiO$_2$ in the dacite field. Both populations contain data points from the two crystals that were analyzed.
6.2 Geothermometry

6.2.1 Titanium-in-Quartz Geothermometer

EPMA line transects were performed across the subhedral quartz crystals found in BB04 and BB24 in order to obtain titanium concentrations. EPMA data for the titanium-in-quartz geothermometer can be found in Appendix E. The transects reveal titanium zonation with Ti varying from 154 ppm to 52 ppm in BB04 and 104 ppm to 58 ppm in BB24 (Figure 6.06). There does not appear to be systematic variation of Ti concentration in either profile nor similarities in variation between the profiles. The highest concentration of titanium in BB04 can be found at 425 µm from the core of the crystal.
Figure 6.06. Titanium-in-quartz plots for quartz from BB04 and BB24. EPMA transects from the core to rim of two quartz crystals from the Banco Bonito Flow. The plots show the titanium concentration in ppm, the calculated titanium-in-quartz temperatures at titanium activities from 0.9 to 0.5, and the calculated titanium-in-quartz temperatures at a titanium activity of 0.6.

which is approximately 875 µm in radius. The lowest concentration can be found at 75 µm from the core of the crystal. The highest titanium concentration in BB24 can be found at two points: 275 µm and 725 µm from the core of the crystal, which is approximately 750 µm in radius. The lowest concentration can be found at 425 µm from the core.
The titanium-in-quartz geothermometry equation requires an estimate of the activity of TiO₂ in the magma (Wark and Watson, 2006). This value can potentially range from 1, in which the system is saturated in titanium, to 0, where there is no titanium in the system. No rutile (or other such titanium-bearing mineral) was observed in any of the units. Thus, it is highly unlikely that the TiO₂ activity in the system is 1. The average rhyolite usually has an activity of approximately 0.6.

Assuming a TiO₂ activity of 0.6 in the system, the average temperatures for BB04 and BB24 would be 801 ± 34 °C and 774 ± 19 °C, respectively. At that same activity, the highest and lowest calculated temperatures for BB04 would be 874 °C and 730 °C. Likewise for BB24, the highest and lowest temperatures would be 817 °C and 743 °C. For both crystals, apparent fluctuations of approximately 140 °C are recorded.

6.2.2 Fe-Ti Oxide Geothermometer

The Fe-Ti oxide geothermometer following Ghiorso and Evans (2008) was used to obtain independent temperature measurements in order to confirm the validity of the titanium-in-quartz temperatures. Appendix F contains EPMA data for the oxide crystals. Samples BB04 and BB19 were chosen due to the high abundance of oxide crystals. Since the oxides were small, point analyses were performed on individual crystals. Twelve magnetite and four ilmenite crystals were found in BB19. Fourteen magnetite and no ilmenite crystals were found in BB04. Thus, no temperatures could be calculated for BB04. The results from BB19 ranged from 830 °C to 887 °C (Table 6.01). The average temperature was 848 ± 22 °C.
As discussed in Chapter 3, the age of the EFM has been quite difficult to constrain. In an attempt to obtain an age, this study employed the use of $^{40}$Ar/$^{39}$Ar geochronology on euhedral biotite and pristine groundmass glass from samples BB17 and BB19, respectively. The analytical parameters of and data obtained from these analyses can be found in Appendix G.

### 6.3.1 Biotite

BB17 biotite aliquot #1 produced a generally concordant age spectrum with the exception of an older first step age (Figure 6.07). The total gas age is $621 \pm 10$ ka. In the age spectrum, steps 5 to 10 (64.7\% of the total $^{39}$Ar released) define a slightly younger plateau age of $589 \pm 24$ ka. Steps 3 through 11 produce an isochron age of $478 \pm 27$ ka.

### Table 6.01 Fe-Ti Oxide Geothermometer Temperatures

<table>
<thead>
<tr>
<th>Magnetite</th>
<th>Ilmenite</th>
<th>Temperature °C</th>
</tr>
</thead>
<tbody>
<tr>
<td>BB19 #14</td>
<td>BB19 #16</td>
<td>877</td>
</tr>
<tr>
<td>BB19 #20</td>
<td>BB19 #16</td>
<td>835</td>
</tr>
<tr>
<td>BB19 #11</td>
<td>BB19 #16</td>
<td>833</td>
</tr>
<tr>
<td>BB19 #14</td>
<td>BB19 #13</td>
<td>877</td>
</tr>
<tr>
<td>BB19 #18</td>
<td>BB19 #17</td>
<td>842</td>
</tr>
<tr>
<td>BB19 #18</td>
<td>BB19 #13</td>
<td>833</td>
</tr>
<tr>
<td>BB19 #12</td>
<td>BB19 #13</td>
<td>856</td>
</tr>
<tr>
<td>BB19 #18</td>
<td>BB19 #15</td>
<td>830</td>
</tr>
<tr>
<td>BB19 #20</td>
<td>BB19 #15</td>
<td>832</td>
</tr>
<tr>
<td>BB19 #11</td>
<td>BB19 #15</td>
<td>830</td>
</tr>
<tr>
<td>BB19 #14</td>
<td>BB19 #18</td>
<td>887</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Average Temperature</th>
<th>848.4</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-sigma standard deviation</td>
<td>22.0</td>
</tr>
</tbody>
</table>

Numbers by the sample name denote the crystal used to calculate the temperature.
Figure 6.07. $^{40}$Ar/$^{39}$Ar isochron plot and age spectrum for BB17 Biotite 1.
This isochron age is defined by 82% of the $^{39}$Ar released. The initial argon $^{40}$Ar/$^{36}$Ar ratio was $303.20 \pm 1.8$. Ca/K ratios and radiogenic yields indicate that a pure, unaltered biotite separate was analyzed.

BB17 biotite aliquot #2 produced a generally concordant spectrum, with an old first step age immediately followed by two step ages that are slightly younger than the following plateau (Figure 6.08). The total gas age is $651 \pm 13$ ka. A plateau age of $615 \pm 20$ ka which is defined by steps 4 through 10 (82% of the total $^{39}$Ar released) in the age spectrum. This sample also gave an isochron age of $575 \pm 15$ ka (Figure 6.08). The isochron is defined by 92% of the $^{39}$Ar released from steps 4 through 12. The initial argon $^{40}$Ar/$^{36}$Ar ratio was $299.6 \pm 1.6$. Ca/K ratios and radiogenic yields indicate that a pure, unaltered biotite separate was analyzed.

6.3.2 Groundmass Glass

BB19 glass aliquot #1 yields neither an isochron nor a plateau age. The total gas age is $125 \pm 1$ ka. The age spectrum is discordant and starts with a very slightly older (with high uncertainty) first step in comparison with the subsequent steps (Figure 6.09a). Steps 2 through 10 vary above and below approximately 120 ka. The final two steps are considerably older than the previous steps. Ca/K ratios and radiogenic yields indicate that a pure, unaltered sample of glass was analyzed.

BB19 glass batch #2 also did not yield an isochron or a plateau. The total gas age was $129.82 \pm 0.80$ ka. The age spectrum is very discordant with an old first step and the subsequent steps shallow out to a gentle saddle shape (Figure 6.09b). Steps 2 through 4 are progressively younger while steps 6 through 13 are progressively older. With the
Figure 6.08. $^{40}\text{Ar}/^{39}\text{Ar}$ isochron plot and age spectrum for BB17 Biotite 2.
Figure 6.09. $^{40}$Ar/$^{39}$Ar age spectra for BB19 glass.
exception of the first and last two steps, the ages are between approximately 100 to 120 ka. The second to last step is notably older than the intermediate steps and is followed by an extremely old final step. Ca/K ratios and radiogenic yields indicate that a pure, unaltered glass separate was analyzed.

6.4 Thermal Ionization Mass Spectrometry

TIMS analyses were chosen for representative samples ECH01, BR01, and BB01. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are 0.512459, 0.512444, and 0.512443 for samples ECH01, BR01, and BB01, respectively. The $\varepsilon_{\text{Nd}}$ values are -3.5, -3.8, and -3.8 for ECH01, BR01, and BB01, respectively. The ratios of $^{87}\text{Sr}/^{86}\text{Sr}$ for ECH01, BR01, and BB01 are 0.704795, 0.704639, and 0.704686, respectively. The corrected $^{208}\text{Pb}/^{204}\text{Pb}$ ratio values are 37.57, 37.51, and 37.52 for ECH01, BR01, and BB01 respectively. The corrected $^{207}\text{Pb}/^{204}\text{Pb}$ ratio values for ECH01, BR01, and BB01 are 15.49, 15.48, and 15.48, respectively. The corrected $^{206}\text{Pb}/^{204}\text{Pb}$ ratio values for ECH01, BR01, and BB01 are 17.66, 17.65, and 17.64, respectively. Reported uncertainties are 2-sigma of mean. Sr, Nd, and Pb isotopic data can be found in Appendix H.

On a plot of $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $\varepsilon_{\text{Nd}}$, the samples fall between the various basalts that have been erupted within and in close proximity to the Rio Grande Rift and the Proterozoic lower crust of the southwestern United States (Figure 6.10). The samples plot within the enriched quadrant. The EFM units are generally homogeneous with respect to isotopic compositions.
6.5 X-ray Fluorescence and Inductively Coupled Plasma Mass Spectrometry

6.5.1 Major Element Geochemistry

All analyzed samples plot as low silica rhyolites (~73 to 75% SiO₂) on the Le Bas et al. (1986) classification diagram (Figure 6.11). The ECPB and BRI samples plot in a distinct cluster that has slightly lower SiO₂ values than the BBF samples. The SiO₂
average for the ECPB and BRI samples is 72.78 whereas the BBF is 73.71. Appendix I contains the XRF major element data.

On diagrams of the major elements versus SiO$_2$ (Harker variation diagrams), the three units usually plot in distinct clusters (Figure 6.12). On most plots, the ECPB and the BRI samples form a distinct group from the BBF samples. General trends are evident between major elements and increasing SiO$_2$. A nearly linear decrease is observed in Al$_2$O$_3$. A general decrease is shown by FeO, MgO, and P$_2$O$_5$. There is a general increase in both Na$_2$O and K$_2$O. The ECPB and BRI samples do not fall within uncertainty of one another in the FeO diagram. The data for CaO and TiO$_2$ are not shown nor discussed.
Figure 6.12. Major element Harker variation diagrams of the East Fork Member. Two-sigma uncertainty is shown by the error bars in the upper right corner of each chart. CaO and TiO are omitted due to the very high two-sigma uncertainty. All oxides are in weight percent.

because the uncertainties associated with the individual analyses are larger than the observed variability. For all other major elements, the variability between the samples is greater than analytical uncertainty.
6.5.2 Trace Element Geochemistry

Incompatible and compatible trace elements were plotted because they are most likely to show chemical evolution within the magma chamber (e.g., fractional crystallization). To plot the trace elements, an incompatible, non-mobile trace element should be used on the x-axis. Trace elements that have good analytical precision and are highly incompatible include Cs, Rb, and Nb. Nb was chosen to represent the x-axis variable because it is the least mobile. All trace element data can be found in Appendices I and J.

The plots of Nb versus trace elements can be divided into groups based on their increasing or decreasing trends with increasing Nb (Figures 6.13a and 6.13b). The following show a general increase, in which data points are not well-correlated: Y, Rb, Ce, Nd, and Lu. A nearly linear increase, in which the data points are well-correlated, is seen in the Ta, Th, Yb, and U plots. A nearly linear decrease is seen in the Eu plot. The following show a general decrease: Sr, Zn, and Ba.

The BBF samples have the highest concentrations of incompatible trace elements and the lowest concentrations of compatible elements. The BRI sample has the lowest concentrations of incompatible elements and has the highest concentrations of compatible elements. The ECPB samples have intermediate amounts of both compatible and incompatible trace elements. Zn and Ba are the exceptions to these observations. The BRI sample contains less Zn than both the ECPB and the BBF samples. The BRI contains less Ba than the ECPB samples.
Figure 6.13a. Plots of Nb versus trace elements measured by ICPMS. The two sigma uncertainty is smaller than the symbols shown. All trace elements were measured by ICPMS and are in ppm.
Figure 6.13b. Plots of Nb versus trace elements measured XRF. The two-sigma uncertainty is shown by the error bars in the upper corner of the plots. All trace elements were measured by XRF and are in ppm.
6.5.3 Ratios of Incompatible Trace Elements

Ratios of similarly incompatible trace element are unaffected by upper crustal processes such as fractional crystallization. Thus, the relative proportions of the trace elements may represent the initial composition of the magma. Again, Nb was chosen as the x-axis variable. In general, all of the units cluster close together (Figure 6.14). All of the samples vary within a span of 0.07 for Ta/Yb and 0.04 for Th/Nb. In the Y/Yb and Th/Yb plots, all of the samples vary within a span of 0.4.

Figure 6.14. Plots of Nb versus ratios of incompatible trace elements. The two-sigma uncertainty is smaller than the symbols shown. The trace elements are in ppm.

6.5.4 REE and Primitive Mantle Spider Diagrams

A rare earth element (REE) plot displays the REE arranged from lightest to heaviest versus normalized rock/chondrite values on a logarithmic scale. All three of the
EFM units exhibit the same pattern and are nearly indistinguishable from one another (Figure 6.15). In general, there is a fairly steep negative slope in the LREE and a shallow to slightly positive slope in the HREE. All samples exhibit a small Eu anomaly.

To further explore the geochemical similarities and differences between the ECPB, the BRI, and the BBF, a primitive mantle normalized spider diagram was plotted (Figure 6.16). The foremost feature of the diagrams is the very large positive anomaly in Pb content. There is also a positive anomaly in U and Th. The notable negative anomalies include: Ba, Nb, Sr, and Eu. The Eu anomaly, in particular, is rather shallow. For the Sr anomaly, the BRI sample plots slightly higher than the ECPB and the BBF samples.

Figure 6.15. Normalized rare earth element diagram of the East Fork Member. Chondrite values from Sun and McDonough (1989).
Figure 6.16. Normalized spider diagram of the East Fork Member. Primitive mantle values from Sun and McDonough (1989).
CHAPTER 7

INTERPRETATIONS AND DISCUSSION OF MODELS FOR THE PRODUCTION AND EVOLUTION OF THE EAST FORK MEMBER MAGMA

In this chapter the magma production models introduced in Chapter 1 will be considered in light of the data discussed in the previous chapter. The data may help elucidate if the model is applicable to the EFM. After consideration of each model, a comprehensive model from the production to eruption of the EFM magma is presented.

7.1 Interpretations of Petrography

7.1.1 Crystal Sources and Histories

Textures indicative of open-system processes are present in most of the large crystals. These textures include resorption, rim overgrowths, and reaction rims. All these textures could indicate crustal melting and/or magma mingling may have taken place. An intruding mafic magma could induce crustal melting. If plagioclase, biotite, and amphibole were derived from a silicic crustal rock, an increase in temperature could have caused some of the crystals to resorb (Figures 5.01, 5.02, 5.04, 5.05). An increase in temperature could also have caused some of the biotite to react and form amphibole overgrowths (Figure 5.03). The morphology of the resorbed crystal aggregates suggests that they may have been derived from a pluton due to the interlocking nature of the crystals (Figure 5.04). However, if the crystal aggregates were derived from a metamorphic rock (i.e. gneiss), it would be expected that they show indicators of strain in the crystals (e.g., undulose extinction), which they do not.

Some textures could indicate a change in magma chemistry. The reaction rim of clinopyroxene around the olivine crystal in sample ECF01 suggests the olivine became
exposed to a more silicic magma. All of the rim overgrowths on plagioclase cores may be indicators of a change in magma chemistry and/or temperature (Figures 6.02, 6.03). Reverse, normal, and patchy zonation is observed in the plagioclase (Chapter 6). Several studies have used major element zonation in plagioclase to constrain magma chamber dynamics. Large fluctuations in anorthite content are generally interpreted to reflect major thermal and compositional changes in the magma chamber (e.g., Anderson, 1983; Nixon and Pearce, 1987; Pearce et al., 1987; Pearce and Kolisnik, 1990; Singer et al., 1995; Tepley et al., 1999; Ginibre et al., 2002). In contrast, small fluctuations may be simply the result of convection (e.g., Lofgren, 1974; Lofgren and Gooley, 1977; Tsuychiyama, 1985; Lasaga, 1982; Loomis, 1982; L’Heureux and Fowler, 1994).

Normal zonation typically occurs during fractional crystallization. Reverse zonation may take place when calcium is introduced to the system, typically by injection of a mafic magma and a subsequent increase in temperature (e.g., Hibbard, 1991; Troll and Schminke, 2001; Ginibre et al., 2002, 2004). The abrupt and large fluctuations in anorthite content seen in the EFM samples suggest that changes in magma chemistry and/or temperature occurred.

The chemistry of the brown glass inclusions in the resorbed plagioclase and high Mg numbers of the biotite and amphibole are consistent with derivation from an andesitic source (Figure 6.05). If the andesitic source had originated from a mantle-derived mafic magma, it would be expected that the isotopic values of the EFM would indicate the presence of mantle material (discussed below).

When compared to the accepted ages of the EFM eruptions at approximately 55 ka and 40 ka (Goff and Gardner, 2004), the biotite $^{40}\text{Ar} / ^{39}\text{Ar}$ isochron ages and age
spectra are almost an order of magnitude older in apparent age. The large euhedral biotite crystals in BB17 gave isochron ages of $478 \pm 27$ ka and $575 \pm 15$ ka (Figures 6.07 and 6.08). Since the biotite did not produce geologically reasonable $^{40}$Ar/$^{39}$Ar ages, it is possible that these crystals are xenocrystic.

Therefore, based on the observations and data discussed above, it may be possible that the resorbed and euhedral large crystals are a mixture of restite and xenocrysts. This interpretation is also supported by the studies of Self et al. (1988) and Wolff and Gardner (1995). In contrast, the small euhedral crystals show no textural open-system indicators and, thus, may be juvenile phenocrysts.

Other studies have documented mixed mafic-silicic crystal populations. Dacite lavas near Clear Lake, California contain bimodal phenocryst populations which exhibited open-system textures (Stimac and Pearce, 1992). Interaction between earlier formed silicic magma bodies and a basaltic andesite magma is thought to have produced this mixed population. The silicic crystals typically displayed resorption, compositional reversals, a partial reaction zone progressing inward from the crystal margins, and coronas that formed by diffusion-limited reactions in dissolution boundary layers. Sodic plagioclase, in particular, showed dissolution, reverse zonation at crystal rims, and fritted texture. Calcic plagioclase thought to be derived from the basaltic andesite magma exhibited normal zonation at crystal rims. Similar textures are observed in plagioclase the EFM, thus similar processes may have been involved in the crystals’ history.
7.1.2 Plagioclase Source Models

The heterogeneity in composition of the plagioclase cores in the EFM is indicative of multiple sources. In general, the plagioclase cores comprise two populations: sodic plagioclase cores and calcic plagioclase cores (Figures 6.02 and 6.03). Thus, two distinct plagioclase sources are implied. The andesitic brown glass inclusions in the plagioclase suggest that a mafic magma may be involved in the derivation of the more calcic cores (Figures 5.02, 5.05, and 6.05). A mafic magma could provide the thermal input required to partially melt silicic crustal rocks. However, a mafic magma would probably not have enough thermal energy to melt a mafic pluton. Thus, the oligoclase cores could be derived from the melting of a silicic crustal source rock (i.e. the cores are restitic). The andesine cores could have been derived from the mafic magma.

A similar mixed plagioclase population was documented in volcanic rocks erupted from Unzen volcano in Japan (Browne et al., 2005). Resorption and rim overgrowths were present in both populations. Plagioclase chemistry suggested derivation from both dacite and basalt. Browne et al. (2005) suggested that the Unzen eruptions were triggered by basalt injecting into a dacitic magma chamber. When basalt was injected into the chamber, the dacitic plagioclase were engulfed by the basalt and resorbed. Basaltic plagioclase were exchanged to the dacitic magma by enclaves that disaggregated. Similar to the plagioclase exchange at Unzen, enclave disaggregation could explain the exchange of mafic crystals to the silicic melt in the EFM.

In contrast to the cores, the rim overgrowths on plagioclase in the EFM record a simple crystallization history in a relatively homogeneous magma body. The rim compositions are similar for both the reversely zoned and normally zoned plagioclase
If the rim overgrowths crystallized in commingling melts, the composition of the rims would have varied. Complex zonation (i.e. large compositional amplitudes, dissolution surfaces, and multiple composition changes) is documented in plagioclase rim overgrowths in their study of Ignimbrite “A” of the Canary Islands (Troll and Schmincke, 2002). No complex zonation is evident in the EFM plagioclase rims. Thus, such a complex crystallization history is not considered to be applicable to the EFM plagioclase overgrowths.

7.2 Diffusive Loss of Argon from Biotite

Based on the crystal morphology, \(^{40}\text{Ar}/^{39}\text{Ar}\) ages older than the accepted age range, and the high Mg numbers, the large euhedral biotite in the BBF could have two possible sources: a mafic pluton or a intruding mafic magma. To test the hypothesis that the crystals were derived from a partially melted intermediate to mafic intrusive body, an equation to model diffusion from a cylinder was employed to calculate the diffusive loss of argon from a biotite crystal (Crank, 1975). The biotite diffusion parameters of Harrison et al. (1985) were used in the calculation (Appendix K). Several basic assumptions were made to use this equation. These assumptions include that argon behaves as a highly incompatible trace element, argon diffusion is compositionally dependent in biotite, and rhyolite has high argon solubility. Since the EFM biotite show little to no Fe and Mg zonation, diffusion of argon from the crystal would not be hindered by compositional zonation. Thus, if a biotite crystal derived from a partially melted pluton resided in a magma, any radiogenic argon that had accumulated would completely diffuse out of the biotite crystal given enough time and sufficient temperature. Using the
average magma temperatures obtained by the titanium-in-quartz and Fe-Ti oxide geothermometers (801, 774, and 848°C); the following times (in years) are the maximum crystal residence time allowed before total argon loss: 0.12, 0.22, and 0.05, respectively. If the crystals were in the magma for longer than the maximum residence time, they would have been completely reset and be expected to give an eruption age matching those of the accepted ages. Given the very short amount of time that the biotite could be in the magma without completely resetting, it is considered unlikely that they are from a partially melted intermediate to mafic pluton.

As stated above, another possibility is that the biotite was derived from a mantle-sourced mafic magma. Magmas derived from the mantle can contain excess $^{40}$Ar (Kelley, 2002). If a silicic magma interacted with the mantle-derived magma, biotite and excess $^{40}$Ar could be transferred to the silicic magma. The excess $^{40}$Ar could remain trapped in the mantle-derived biotite crystals since they likely would not resorb when exposed to a cooler silicic magma. The accumulation of radiogenic $^{40}$Ar produced from the decay of $^{40}$K would take place as usual once the magma erupted. Thus, the older apparent $^{40}$Ar/$^{39}$Ar biotite ages of the EFM could be due to excess $^{40}$Ar (Figures 6.07, 6.08). The saddle-shaped age spectra for groundmass glass from the BBF indicate that there may be excess $^{40}$Ar present (Figure 6.09). Therefore, it is possible that the large euhedral biotite were derived from a mantle-derived magma, and they, as well as the EFM magma itself, contain a small amount of excess $^{40}$Ar reflecting this mantle-source history.
7.3 Isotopic Mixing Models

The radiogenic isotopes could elucidate if the EFM is result of the direct fractional crystallization of basalt, a pure crustal melt, or a mixture between isotopic reservoirs. If the magma is a result of the mixing of two isotopic reservoirs, an isotopic mixing model may constrain the proportions of the two sources.

The continental crust beneath the Jemez mountains volcanic field (JMVF) is approximately 1.7 Ga (Van Schmus and Bickford, 1983). Thus, the isotopic composition of the crust is very different than that of the underlying mantle. The compositions of the upper and lower crust are likely to be different from one another as well (DePaolo and Wasserburg, 1976). The isotopic compositions of the units of the EFM are distinct from mid-ocean ridge basalts and the Precambrian crust beneath the JMVF (Figure 6.10). Therefore, an isotopic mixing model has been attempted in order to constrain the percentage of mantle-derived material and crustal component needed to produce the Sr and Nd isotopic compositions seen in the EFM.

7.3.1 Isotopic Mixing Model Using Spell et al. (1993) Data

To model isotopic mixing, end member compositions must be chosen. Spell et al. (1993) compiled available data and chose three end members: the mantle, Proterozoic lower crust, and a hypothetical hybridized lower crust. These three end members are used in this thesis due to their applicability to the post-collapse units in the Valles caldera. The mantle end member values are $\varepsilon_{\text{Nd}} = +7$, Nd = 30 ppm, $^{87}\text{Sr}/^{86}\text{Sr} = 0.7030$, and Sr = 1000 ppm. These values were derived from the various basalts that have been erupted in close proximity to and within the Rio Grande Rift. Two lower-crustal end members were chosen. No upper-crustal end member was chosen since the mantle-
derived magma would first inject into, and possibly melt, the lower crust. Lower-crustal xenoliths and exposed Proterozoic granitic rocks give estimates as to what the lower-crustal composition might be. Spell et al. (1993) designated values of $\varepsilon_{\text{Nd}}$ of -12, Nd = 30 ppm, $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ to 0.710, and Sr = 500 ppm for unmodified lower crust. However, the lower crust beneath the JMVF has likely been injected by basalt for the past 16 Ma (Gardner and Goff, 1984), thus it is possible that the lower-crust is now hybridized. The hybridized lower-crustal end member values used in the model are $\varepsilon_{\text{Nd}} = -8$, Nd = 30 ppm, $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$ to 0.710, and Sr = 500 ppm (Spell et al., 1993). The isotopic data for all three units of the EFM fall on the mixing line between the hybridized lower-crustal and mantle end members (Figure 7.01). There is an approximate 75% hybridized lower-crustal component and 25% mantle component according to this model. If the lower crust is indeed hybridized, then the model result of the mantle component would be an underestimate.

### 7.3.2 Isotopic Mixing Model Using Wolff et al. (2005) Data

Other studies have invoked the use of isotopic mixing models to characterize Jemez volcanic field lavas (Wolff et al., 2005; Rowe et al., 2007). Wolff et al. (2005) compiled isotopic data from Proterozoic metamorphic rocks of the Taos Range, lithic inclusions in the Otowi member of the Bandelier Tuff, and mafic lava flows in the Jemez volcanic field. The $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ for the mafic lavas range from 0.51244 to 0.51273 and 0.7041 to 0.7048, respectively. Chemical characteristics of the mafic magmas show that they are crustally contaminated (Wolff et al., 2000, 2005; Rowe et al., 2007).

Isotopic data from Wolff et al. (2005) was used in three mixing models. The EFM samples fall on mixing lines between samples I-6-CdR (Cerros del Rio Formation) and
Figure 7.01. Isotopic mixing models using end members from Spell et al. (1993). A radiogenic isotope mixing curve with tick marks in 20% intervals from the mantle end member to the hybridized lower crustal end members. MORB = Mid-ocean ridge basalts. MT = Mount Taylor basalts (Perry et al., 1990). RGR = Rio Grande Rift basalts (Perry et al., 1987). ZB = Zuni Bandera basalts (Menzies et al., 1991). The field for the Jemez basalts is from the data of DePaolo and Wasserburg (1976), Perry et al. (1987), and Loeffler et al. (1988). The lower crust field is from southwestern U.S. xenolith data from James et al. (1980), De Paolo (1981), Esperança et al. (1988), Padovani and Reid (1989), and Kempton et al. (1990). The figure is modified from Spell et al. (1993).

T-177 (Taos Range Amphibolite Gneiss) as well as MR 00-9 (Lobato Formation) and T-178 (Taos Range Amphibolite). See Figures 7.02A and 7.02B. According to both models, an approximate 40% mantle component and 60% crustal component is present in the EFM. A mixing line between samples JM93262 (Palizia Canyon) and T-177 (Taos Range Amphibolite) shows an approximate 30% mantle component and 70% crustal component (Figure 7.02C).
Figure 7.02. Isotopic mixing models for the East Fork Member using end members from Wolff et al. (2005). Radiogenic isotope mixing curves with tick marks in 20% intervals between samples: A) I-6-CdR to Amphibolite178. B) MR00-9 to Amphibolite178. C) JM93262 to Amphibolite177. Data from Wolff et al. (2005)
To model isotopic mixing between the mantle and crust, Rowe et al. (2007) used sample E6-8B (Cerros del Rio Formation) as the mantle representative and CCL-1 (lithic in the Otowi) as the crustal representative. The EFM samples do not fall on the mixing line produced by this model.

### 7.3.3 Discussion of Isotopic Mixing Models

In comparison to the isotopic mixing model produced using the Spell et al. (1993) end members, the Wolff et al. (2005) mixing models indicate a greater amount of mantle in the EFM. However, it should be noted that the Wolff et al. (2005) mafic samples contain less Nd and the crustal samples contain more Sr than the respective Spell et al. (1993) end members. It should also be noted that the Spell et al. (1993) hybridized lower crustal end-member assumes contamination of the crust by injected mafic magma. The Wolff et al. (2005) mafic samples are assumed to be crustally contaminated. Therefore, the Spell et al. (1993) and the Wolff et al. (2005) results are underestimates of the mantle and crustal components, respectively, and provide the lower limits of both components.

Evidence for isotopic mixing between the mantle and lower crust has been documented by Karsli et al. (2008) in the Quaternary Erzincan Volcanics in Turkey. Like the EFM, the rhyolites in the Erzincan volcanics have low $^{87}$Sr/$^{86}$Sr ratio values and show a large lower-crustal isotopic signature.

It is common for magmas emplaced in the upper crust to assimilate upper-crustal material. If the EFM assimilated a significant amount of an upper crustal component, the samples would likely have a higher ratio of $^{87}$Sr/$^{86}$Sr and a lower $\varepsilon_{Nd}$ value. Substantial upper-crustal assimilation would pull the EFM samples off the mixing curves between
the mantle and lower crust end members and move them towards the upper-crustal field, adding a third component to the magma. Upper crustal assimilation has been documented in some of the other post-collapse eruptions in the Valles (Spell et al., 1993). However, it is not considered likely that upper-crustal assimilation took place because the isotopic mixing model does not indicate an upper-crustal component is present, and a simple fractional crystallization model explains the trace element data trends (discussed below).

7.4 Fractional Crystallization Model

As discussed in Chapter 6, nearly linear increasing and decreasing trends are evident in plots of selected trace elements vs Nb (Figures 6.13a and 6.13b). In those plots, typically sample BR01 is the least evolved, and sample BB04 is the most evolved. Thus, to better constrain relationships between the three units of the EFM, a fractional crystallization model was tested. Appendix L contains all fractional crystallization model data.

Fractional crystallization calculations were completed using ICPMS trace element data from representative samples BR01 as the parental composition and BB04 as the daughter composition, measured modal abundances from point count data, and representative rhyolite partition coefficients obtained from the GERM website (http://earthref.org/KDD/). However, none of the modal abundances calculated from the point counts produced fractional crystallization models that fit the trace element data. As previously discussed in section 7.1, it is possible that many of the large crystals are restitic or xenocrystic. Thus, the point count modal abundances may not accurately
portray the relative abundances of juvenile phenocrysts crystallizing from the magma. Thus, a model modal abundance was found after several iterations that produced a fit for most of the trace element data at 17.5% fractional crystallization (Table 7.01). Figure 7.03 shows the model fractional crystallization trends for selected trace elements. This model suggests that fractional crystallization was the last process to take place since subsequent open-system processes such as crustal assimilation and/or magma mixing would be expected to cause the model to fail.

Major element data also suggests that fractional crystallization was taking place. Slight decreasing trends in MgO, P$_2$O$_5$, and FeO; a nearly linear decrease in Al$_2$O$_3$; and slight increasing trends in Na$_2$O and K$_2$O are consistent with fractional crystallization (Figure 6.12).
Table 7.01 Model for fractional crystallization of Battleship Rock magma to produce Banco Bonito Flow magma.

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<th>Mineral Phases (modal abundance)</th>
<th>Quartz</th>
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<th>Sanidine</th>
<th>Biotite</th>
<th>HBL</th>
<th>CPX</th>
<th>OPX</th>
<th>Magnetite</th>
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<th>Daughter (Obs.) BB04</th>
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Note: HBL = hornblende, CPX = clinopyroxene, OPX = orthopyroxene
FC = fractional crystallization
Figure 7.03. Fractional crystallization trends of Nb versus selected trace elements. The tick marks are in 5% intervals. The two-sigma uncertainty is smaller than the symbols shown. All trace elements are in ppm.
7.5 Models for Silicic Magmatism

As discussed in Chapter 1, several models have been put forth to explain silicic magma production, especially with reference to caldera-forming systems. The models not only focus on the caldera-forming eruptions, but also on the post-collapse eruptions.

7.5.1 Direct Fractional Crystallization of Basalt Model

If the EFM magma was produced by fractional crystallization of basalt, the radiogenic isotopes should reflect the values of mantle-derived basalt; however, as discussed in section 7.3, they do not. The heterogeneity of the crystal assemblages is also not consistent with simple fractional crystallization. Resorption, presence of crystal aggregates, and reverse zoning seen in the plagioclase crystals would not be easily predicted by this simple model. Thus, the EFM is not considered to be the product of fractional crystallization of basalt.

7.5.2 Pure Crustal Melting Model

Likewise, if the EFM magma was derived only from crustal melt, the radiogenic isotopic values would be expected to reflect solely those of the upper or lower crust beneath the Jemez mountains volcanic field, which they do not. The heterogeneity of the crystal assemblage (i.e. mafic-silicic crystals) and morphology (e.g., reaction rims and rim overgrowths) in the EFM also argues against the applicability of this model. The diffusive loss of argon modeling suggests that the melting of a mafic pluton is not consistent with the $^{40}\text{Ar}/^{39}\text{Ar}$ biotite geochronology results. The mafic glass inclusions would not be predicted in this model. The EFM is, therefore, not considered to be the product of pure crustal melting.
7.5.3 The Long-lived Magma Chamber Model

This model describes the evolution of a single, long-lived upper crustal magma chamber, which is thermally sustained in a liquid state by underlying mafic sills (e.g., Smith, 1979; Hildreth, 1981; Lipman, 1984). Periodic large-scale explosive to small-scale effusive eruptions would tap the same magma reservoir. Thus, eruptions from a single caldera system should be cogenetic if they all are sourced from the same evolving magma chamber. This aspect of the model is not supported since the EFM is distinct from the other post-collapse rhyolites (discussed below).

The results of this thesis demonstrate that the EFM magma is relatively low in SiO₂ and incompatible trace elements (Rb, Cs, HREE, Pb, Th) and has high values of TiO₂, FeO, MgO, CaO and compatible trace elements (Sc, Cr, Sr, Zr, Ba, Eu) in comparison with the other post-collapse rhyolites. Similar trace element compositions for the EFM have been shown in previous studies (Self et al. 1988; Spell et al., 1993), which concluded that in comparison with the other post-collapse rhyolites and the Bandelier Tuff, the EFM is least evolved and compositionally distinct. More detailed data from this thesis supports this conclusion.

The results from this study also demonstrate that the εNd values of the EFM range from -3.5 to -3.8, and the $^{87}$Sr/$^{86}$Sr values range from 0.704795 to 0.704686 (Chapter 6). The εNd values and the initial $^{87}$Sr/$^{86}$Sr values of other post-collapse rhyolites range from -4.6 to -3.7 and from 0.71307 to 0.70512, respectively (Spell et al., 1993). Thus, the EFM has overlapping to slightly higher εNd and lower $^{87}$Sr/$^{86}$Sr values in comparison to the other post-collapse rhyolites. The $^{87}$Sr/$^{86}$Sr values of sanidine from the Otowi
Member of the Bandelier Tuff are 0.7074 to 0.7052 (Wolff and Ramos, 2003). Thus, the EFM is isotopically distinct from the other post-collapse units and the Bandelier Tuff.

The EFM is sparsely porphyritic to porphyritic. Based on point counts, ECH01, BR01, and BB01 contain 6.5%, 2.0%, and 14.4% crystals, respectively (Table 5.01). Spell et al. (1993) discuss the petrology of the post-collapse rhyolites and show that the eruptive units range from aphyric to porphyritic and contain relatively different crystal assemblages and abundances from one another. In comparison with the crystal assemblages of the other post-collapse units, the EFM is generally more crystal-poor and lacks an abundance of sanidine and quartz (Table 5.01).

The comprehensive results of this thesis are consistent with the observations made by previous studies mentioned above. Therefore, the EFM is indeed chemically, isotopically, and petrographically distinct from other post-collapse rhyolites. This conclusion is not consistent with the general model of a long-lived, convecting magma chamber, as this model would predict that all post-collapse eruptions should be derived from a single evolving magma chamber.

The other petrographic observations made in this study are also not consistent with the long-lived magma chamber model. Resorption, reaction rims, and other open-system indicators would not be easily predicted if the EFM magma came from the most evolved upper levels of such a magma reservoir. The heterogeneity in crystal size and chemistry does not indicate that the EFM crystals are from a single source. The two different plagioclase core compositions would be difficult to predict using this model. The high Mg numbers for the large euhedral biotite and amphibole in the BBF indicate
that those crystals did not form in a silicic magma. It would be difficult to predict the olivine crystal in the ECPB and the mafic glass inclusions in the resorbed plagioclase using this model. Due to the many inconsistencies of the results of this thesis with the predicted outcomes, the long-lived, large magma chamber model is not considered to be applicable to the EFM.

7.5.4 The Crystal Mush Model

In this model, rhyolite magma is produced by the silicic liquid that collects at the top of the magma chamber due to crystal settling and compaction (Bachmann and Bergantz, 2004, 2008). The crystal mush model suggests that sequential small-scale eruptions should be cogenetic since they tap the same reservoir (Bachmann and Bergantz, 2004). As was already discussed, the EFM is distinct from the other post-collapse rhyolites. This model predicts that the erupted magmas would be high silica rhyolite and aphyric. However, the EFM is low silica and sparsely porphyritic to porphyritic. The heterogeneity of crystal morphology and chemistry observed in the EFM is not consistent with the crystal mush model. The presence of mafic glass inclusions in resorbed plagioclase, high-Mg biotite and amphibole, resorption, and rim overgrowths would generally not be predicted. Thus, the crystal mush model is not considered to be applicable to the EFM.

7.5.5 The Rapid Production and Eruption Model

The Huppert and Sparks (1988) model states that rhyolitic melt can be rapidly produced by the injection of a mafic magma into the crust, which undergoes partial melting. Simultaneous melting and crystallization may take place in the source region. The melting takes place in the partial melting region (Figure 7.04). Resorption takes
Figure 7.04. Schematic diagram of the roof region in the rapid magma production model. A) Partial melting region: Partial melting of the crust takes place here. B) Thermal boundary layer: Resorption of crystals takes place here. Local areas of instability create plumes that descend into the convecting interior. C) Convecting interior: Crystallization takes place here. Plagioclase cores can become nuclei for new growth. Modified from Huppert and Sparks (1988).

place in the thermal boundary layer. Crystallization takes place in the cooling, convecting interior. Such simultaneous melting and crystallization could provide an explanation for the variety of observed crystal morphologies and chemistries in the EFM. The mix of restite and juvenile phenocrysts as well as resorption and other open-system indicators seen in the EFM are consistent with this model. Resorbed plagioclase cores becoming the nuclei for new crystal growth is also predicted in this model and consistent with the petrographic observations made in this study. Independent melting and magma production events allows for each batch of erupted magma to be distinct from one another. The EFM is distinct from the other post-collapse rhyolites. Wolff and Gardner (1995) in a less comprehensive study also suggested the Huppert and Sparks model is applicable to the EFM.
Thus, the data from this study best fits the rapid silicic magma production and eruption model put forth by Huppert and Sparks (1988), but with some modification. In their model, a mafic magma provides the heat required to melt the crust which produces a silicic magma but without interaction between the two magmas. The evidence for the interaction and exchange of crystals between a silicic melt and mafic magma seen within the EFM can be explained by a model in which a mafic magma provides the thermal input necessary to partially melt the crust and then interacted with the melt.

7.5.6 The Proposed EFM Model

The EFM magma may have been produced when a mantle-derived basaltic magma became andesitic, possibly by fractional crystallization. The ponding of basalt and subsequent fractional crystallization may have occurred at the base of, or within, the lower crust due to the lower density of the lower crust as compared to basalt (Figure 7.05). Fractional crystallization of basalt could have allowed for the magma to become andesitic in composition. High-Mg biotite and amphibole as well as Ca-rich plagioclase crystallized from the andesitic magma. The andesitic magma rose in the lower crust, partially melted the lower crust, and interacted with this melt, as indicated by the isotopic mixing model. In a similar model put forth by Solano et al. (2012), the crystallization of basalt sills in the lower crust can generate residual melt from the basalt to migrate into and melt the lower crust. The resulting hybrid magma is typically granitic in composition and has chemical signatures of a mixed crustal-mantle origin. Other studies have also
Figure 7.05. Schematic diagram of the model for the petrogenesis of the East Fork Member. A) Basalt ponds at the base of or within the lower crust and undergoes FXL to become andesitic. B) The andesitic magma rises and melts the lower crust. An exchange of crystals takes place. C) The now rhyolitic magma rises into the upper crust, and fractional crystallization takes place. D) The eruptions of the ECPB, BRI, and BBF take place.
documented evidence of basalt becoming andesitic and mixing with rhyolitic magma (Varga et al., 1990). In the case of the EFM, when the andesitic magma interacted with the silicic melt, an exchange of crystals may have taken place. An increase in temperature could cause any crustally-derived minerals (e.g., sodic plagioclase and quartz) to resorb. If the melting of hydrous minerals (e.g., biotite and amphibole) took place, then water once contained in the crystals would be released into the crustal melt. Thus, anhydrous minerals from the andesitic magma (e.g., calcic plagioclase) would resorb once exposed to the water-bearing crustal melt. The amphibole and biotite from the andesitic magma would not resorb, because they were hydrous minerals exchanged to a cooler magma (Section 7.01).

This proposed model for the EFM is consistent with features that suggest that many of the large resorbed crystals have been recycled from a silicic pluton. Similarly, in a study of the Devils Kitchen rhyolite, zircons were shown to be recycled from different magma batches (Miller and Wooden, 2004). The Devils Kitchen rhyolite may have been produced by multiple injections of basalt, thus causing melting of older partially or completely crystallized granitic plutons and subsequent crystallization of the rejuvenated or newly generated melt. The restitic zircons became nuclei for new growth. This model is analogous to the large resorbed feldspar crystals in the EFM.

It is unknown whether the EFM magma became thermally stable and compositionally homogeneous prior to or after rising to the upper crust. Regardless, fractional crystallization (FXL) would be expected to take place once the magma began to cool. The FXL model for trace elements in the EFM indicates that this was the last process to take place prior to eruption. A small amount of FXL took place prior to the
ECPB and BRI eruptions (~55 ka) as indicated by the BRI and ECPB trace element trends (Figures 6.13a and 6.13b). The BBF magma continued fractionating until it erupted approximately 15 ka later. This continued FXL is indicated by the trace element trends (Figures 6.13a and 6.13b) and the greater abundance of small euhedral crystals in the BBF.

The Huppert and Sparks (1988) model has been suggested for other volcanic systems. Perry et al. (1990) propose the model may be applicable to the Mount Taylor volcanic field. The Mount Taylor volcanic field is in northern New Mexico and, like the Jemez mountains volcanic field, overlies the Jemez lineament. Crustal assimilation and fractional crystallization are suggested to be important in the production of Mount Taylor magmas. Many of the rhyolitic and latite magmas were generated by the assimilation of lower crust by basalt, followed by the rise of the magma into the upper crust, and fractional crystallization taking place.

The EFM magma production and eruption was a discrete event when compared to the previous post-collapse eruptions. Similarly, the Long Valley system has been suggested to be characterized by six spatially discrete magma generation events (Hildreth, 2004). These events took place prior to and after the caldera-forming eruption of the Bishop Tuff (760 ka). Each magmatic event was created by basaltic intrusion in the deep crust in response to extensional unloading and decompression melting of the upper mantle. Thus, the upper-crustal magma reservoirs of the Long Valley system are fed by discrete deep crustal melting zones. The Valles caldera system may be similar.

Spell et al (1993) suggested that the post-collapse rhyolites are composed of five discrete events fed by basalt-induced deep crustal melting events and are sourced from
separate magma chambers. The eruptions took place in temporal clusters. Several hundred thousand year lulls in volcanism took place between the clusters of eruptions. A substantial lull in volcanism took place between the eruptions of the South Mountain Member at approximately 520 ka (Spell and Harrison, 1993) and the eruptions of the EFM at 55 ka and 40 ka (Goff and Gardner, 2004). The EFM is compositionally distinct from the previous eruptions. Thus, the eruption of the EFM may represent the onset of a new series of eruptions.
The magmatic and eruptive history of the East Fork Member (EFM) is complex and best explained by the rapid production model of Huppert and Sparks (1988) with slight modification. A mantle-derived basalt ponded at the base of or within the lower crust, underwent fractional crystallization, and became andesitic in composition. The andesitic magma began to crystallize biotite, amphibole, and calcic plagioclase, rose through the crust, triggered partial melting of the lower crust, and mixed with the lower crustal melts. The restitic sodic plagioclase, quartz, and biotite crystals from the crustal melt resorbed due to the change in temperature and magma composition. The resorbed calcic plagioclase as well as the unresorbed biotite and amphibole were derived from the andesitic magma. It is unknown whether the EFM magma became thermally stable and compositionally homogeneous prior to or after rising to the upper crust. Fractional crystallization began to take place once the magma began to cool. Rim overgrowths crystallized over the calcic and sodic plagioclase cores. The small euhedral crystals in the EFM are likely to be juvenile phenocrysts. The EFM magma erupted as the ECPB and BRI at approximately 55 ka. The remaining magma continued to fractionate until it erupted as the BBF at approximately 40 ka.
Appendix A. Sample Locations and Analytical Methods Performed

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## Appendix B. Electron Microprobe Analyses of Feldspar (continued)

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### Appendix B. Electron Microprobe Analyses of Feldspar (continued).

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## Appendix B. Electron Microprobe Analyses of Feldspar (continued).

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Appendix B. Electron Microprobe Analyses of Feldspar (continued).

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### Appendix B. Electron Microprobe Analyses of Feldspar (continued).

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<th>MgO</th>
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### Appendix C. Electron Microprobe Analyses of Biotite and Amphibole in weight percent. Erroneous data not included in results has been omitted.

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### Appendix C. Electron Microprobe Analyses of Biotite and Amphibole (continued).

| Sample       | Distance (µm) | SiO₂   | FeO   | CaO   | Na₂O  | Al₂O₃  | MnO   | K₂O   | MgO   | TiO₂  | Total |
|--------------|---------------|--------|-------|-------|-------|--------|-------|-------|-------|-------|--------|-------|
| BB19 Amph1   | 0             | 48.931 | 13.553| 11.726| 1.431 | 6.758  | 0.624 | 0.722 | 15.651| 1.435 | 100.830|
|              | 50            | 48.519 | 13.596| 11.336| 1.543 | 6.559  | 0.707 | 0.738 | 14.741| 1.507 | 99.246 |
|              | 75            | 48.436 | 14.144| 11.444| 1.326 | 6.439  | 0.666 | 0.677 | 15.088| 1.359 | 99.579 |
|              | 225           | 47.602 | 14.107| 11.432| 1.364 | 6.332  | 0.769 | 0.667 | 15.174| 1.352 | 98.798 |
|              | 250           | 48.886 | 12.018| 13.282| 1.524 | 6.054  | 0.843 | 0.445 | 14.371| 1.373 | 98.796 |
|              | 275           | 48.654 | 13.689| 11.551| 1.282 | 5.897  | 0.835 | 0.695 | 14.850| 1.363 | 98.816 |
|              | 300           | 48.434 | 13.539| 11.362| 1.285 | 6.447  | 0.675 | 0.667 | 15.180| 1.386 | 98.976 |
|              | 325           | 48.720 | 14.143| 11.102| 1.348 | 6.309  | 0.762 | 0.726 | 14.173| 1.305 | 98.589 |
| BB19 Amph2   | 0             | 48.126 | 13.790| 11.304| 1.239 | 6.459  | 0.549 | 0.743 | 15.462| 1.434 | 99.107 |
|              | 50            | 48.528 | 13.892| 11.359| 1.412 | 6.323  | 0.770 | 0.638 | 14.527| 1.350 | 98.800 |
|              | 75            | 48.135 | 13.951| 11.472| 1.459 | 6.394  | 0.721 | 0.742 | 14.643| 1.303 | 98.819 |
|              | 100           | 47.530 | 14.136| 11.290| 1.494 | 6.308  | 0.720 | 0.739 | 14.717| 1.334 | 98.268 |
|              | 125           | 48.712 | 14.120| 11.330| 1.354 | 6.415  | 0.786 | 0.698 | 14.018| 1.358 | 98.791 |
|              | 150           | 49.140 | 13.914| 11.516| 1.267 | 6.198  | 0.765 | 0.718 | 13.970| 1.305 | 98.792 |
|              | 175           | 48.024 | 13.908| 11.548| 1.448 | 6.357  | 0.719 | 0.761 | 14.892| 1.326 | 98.982 |
|              | 200           | 47.969 | 14.050| 11.437| 1.359 | 6.578  | 0.702 | 0.717 | 14.736| 1.429 | 98.977 |
|              | 225           | 48.064 | 13.638| 11.327| 1.358 | 6.497  | 0.711 | 0.716 | 14.114| 1.374 | 97.798 |
|              | 250           | 50.562 | 13.022| 11.504| 1.226 | 6.729  | 0.704 | 0.677 | 14.495| 1.258 | 100.177|
|              | 300           | 52.320 | 8.937 | 20.244| 0.697 | 1.741  | 0.921 | 0.248 | 13.728| 0.224 | 99.060 |
|              | 375           | 48.424 | 14.364| 11.412| 1.361 | 6.636  | 0.735 | 0.765 | 15.034| 1.237 | 99.970 |
|              | 400           | 47.426 | 14.187| 11.168| 1.333 | 6.394  | 0.665 | 0.713 | 14.892| 1.292 | 98.070 |
|              | 425           | 47.574 | 14.254| 11.166| 1.265 | 6.985  | 0.743 | 0.747 | 13.391| 1.431 | 97.554 |
|              | 475           | 49.100 | 14.107| 11.080| 1.365 | 7.066  | 0.736 | 0.788 | 15.033| 1.379 | 100.654|
|              | 525           | 46.612 | 13.753| 11.304| 1.333 | 6.421  | 0.732 | 0.770 | 13.375| 1.318 | 95.617 |
|              | 550           | 46.888 | 13.918| 10.871| 1.471 | 6.277  | 0.694 | 0.730 | 15.280| 1.291 | 97.419 |
|              | 575           | 48.770 | 13.614| 11.329| 1.450 | 6.475  | 0.742 | 0.725 | 14.543| 1.322 | 98.970 |
|              | 600           | 49.165 | 13.002| 10.700| 1.536 | 6.743  | 0.751 | 0.877 | 13.615| 1.298 | 97.687 |
|              | 625           | 46.884 | 13.855| 10.847| 1.328 | 6.031  | 0.726 | 0.718 | 14.400| 1.256 | 96.045 |
### Appendix C. Electron Microprobe Analyses of Biotite and Amphibole (continued).

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<th>CaO</th>
<th>Na$_2$O</th>
<th>Al$_2$O$_3$</th>
<th>MnO</th>
<th>K$_2$O</th>
<th>MgO</th>
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### Appendix D. Electron Microprobe Analyses of Glass Inclusions. Data in weight percent.

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<th>CaO</th>
<th>MgO</th>
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Appendix F. Electron Microprobe Analyses of Oxides in weight percent. Erroneous data not included in results has been omitted.

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<th>MgO</th>
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<th>MnO</th>
<th>TiO₂</th>
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Appendix G. $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology Data.

BB17 Biotite 1.

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<th>$^{37}\text{Ar}$</th>
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<th>rlds</th>
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Cumulative %$^{36}\text{Ar}$ rlds 100.0

note: isotope beams in mV, rld = released, error in age includes J error, all errors 1-sigma

($^{36}\text{Ar}$ through $^{40}\text{Ar}$ are measured beam intensities, corrected for decay for the age calculations)

Total gas age = 621.5 10.03
Plateau age = 589.55 24.27
 Isabel age = 478.00 27.00
(steps 5-10)
(steps 4-11)
Appendix G. $^{40}$Ar/$^{39}$Ar Geochronology Data (continued).

BB17 Biotite 2.

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Cumulative %$^{39}$Ar rlsd = 100.0

Note: isotope beams in mV, rlsd = released, error in age includes J error, all errors 1-sigma
($^{36}$Ar through $^{40}$Ar are measured beam intensities, corrected for decay for the age calculations)
Appendix G. $^{40}$Ar/$^{39}$Ar Geochronology Data (continued).

BB17 Volcanic Glass 1.

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Cumulative %$^{39}$Ar rsl = 100.0

Total gas age = 125.01 ± 1.05

Note: isotope beams in mV, rsl = released, error in age includes J error, all errors 1-sigma

($^{36}$Ar through $^{40}$Ar are measured beam intensities, corrected for decay for the age calculations)
Appendix G. $^{40}$Ar/$^{39}$Ar Geochronology Data (continued).

BB17 Volcanic Glass 2.

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<th>$^{38}$Ar</th>
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Cumulative %$^{39}$Ar rsls = 100.0

Note: isotope beams in mV, rsls = released, error in age includes J error, all errors 1-sigma

($^{36}$Ar through $^{40}$Ar are measured beam intensities, corrected for decay for the age calculations)

Total gas age = 129.82 ± 0.85

No Plateau

No Isochron
## Appendix H. Thermal Ionization Mass Spectrometry Data

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<th>(^{87}\text{Sr}/^{86}\text{Sr}) (measured)</th>
<th>Uncert. 2-sigma</th>
<th>(^{143}\text{Nd}/^{144}\text{Nd}) (measured)</th>
<th>Uncert. 2-sigma</th>
<th>(e_{\text{Nd}}) (calculated)</th>
<th>(^{145}\text{Nd}/^{144}\text{Nd})</th>
<th>(^{208}\text{Pb}/^{204}\text{Pb}) (corrected)</th>
<th>(^{207}\text{Pb}/^{204}\text{Pb}) (corrected)</th>
<th>(^{206}\text{Pb}/^{204}\text{Pb}) (corrected)</th>
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Appendix I. X-ray Fluorescence: Unnormalized Major and Trace Elements

Unnormalized Major Elements (weight percent):

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<th>Na₂O</th>
<th>K₂O</th>
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### Appendix K. Diffusive Loss of Argon in Biotite Data

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**calculated f**

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#### TitaniQ BB24

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#### Fe-Ti Average

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Note: f = 1 is total loss.
Appendix L. Fractional Crystallization Model Data (in ppm)

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Note: Plag = Plagioclase, CPX = clinopyroxene, OPX = orthopyroxene, HBL = hornblende
Appendix L. Fractional Crystallization Model Data (continued)

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Note: F = 0.6 through 0.3 omitted.
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Special Honors and Awards:
Geological Society of America Student Research Grant: Spring 2009

Thesis Title: Petrogenesis of the East Fork Member Rhyolites, Valles Caldera, New Mexico, USA

Thesis Examination Committee:
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Committee Member, Adam Simon, Ph.D.
Committee Member, Michael Wells, Ph.D.
Committee Member, John Wolff, Ph.D.
Committee Faculty Representative, George Rhee, Ph.D.